

Crustal Deformation in Southcentral Alaska: The 1964 Prince William Sound Earthquake Subduction Zone

Steven C. Cohen

Geodynamics Branch, Goddard Space Flight Center, Greenbelt, MD 20771

email: Steven.C.Cohen@.nasa.gov; phone: 301-614-6466

Jeffrey T. Freymueller

Geophysical Institute, University of Alaska, Fairbanks, AK 99775

Outline:

1. Introduction

2. Tectonic, Geologic, and Seismologic Setting

3. Observed Crustal Motion

3.1 Coseismic Crustal Motion

3.2 Preseismic and Interseismic (pre-1964) Crustal Motion

3.3 Postseismic and Interseismic (post-1964) Crustal Motion

4. Models of Southcentral Alaska Preseismic and Postseismic Crustal Deformation

5. Conclusions

6. References

1. Introduction

The $M_w = 9.2$ Prince William Sound (PWS) that struck southcentral Alaska on March 28, 1964, is one of the important earthquakes in history. The importance of this Great Alaska Earthquake lies more in its scientific than societal impact. While the human losses in the PWS earthquake were certainly tragic, the sociological impact of the earthquake was less than that of is earthquakes that have struck heavily populated locales. By contrast Earth science, particularly tectonophysics, seismology, and geodesy, has benefited enormously from studies of this massive earthquake. The early 1960s was a particularly important time for both seismology and tectonophysics. Seismic instrumentation and analysis techniques were undergoing considerable modernization. For example, the VELA UNIFORM program for nuclear test detection resulted in the deployment of the World Wide Standard Seismographic Network in the late 1950s and early 1960s (Lay and Wallace, 1995). Correspondingly, considerable effort was being devoted by the scientific community to understanding seismic source properties, the characteristics of seismic signals, and the interior structure of the Earth. The study of the PWS earthquake benefited from this attention to global as well as local seismological issues and, reciprocally, the study of the PWS earthquake contributed much to the research efforts of the time. In the early 1960's the paradigm of plate tectonics was in its infancy. Many fundamental plate tectonics investigations, such as the study of marine magnetic anomalies by Vine and Mathews (1963), were just beginning to be reported in the scientific literature. Certainly the structure of subduction zones and their overall importance in the scheme of seismotectonics were barely appreciated. However, fundamental investigations into the relationship between seismic and tectonic phenomena were underway and culminated in a surge in seismotectonic findings in the mid and late 1960s and early 1970s. These early studies of the PWS earthquake led credence to the idea that strain was

accumulated and released across low dip-angle megathrusts at plate boundaries such as the one in southcentral Alaska and the Aleutians.

Crustal deformation is an important, measurable phenomenon related to the cycle of stress accumulation and release at tectonic plate boundaries. Thus, crustal deformation measurements, along with seismological and geological investigations, have provided critical information for understanding earthquake processes. In southcentral Alaska geodetic observations span the coseismic epic and a postseismic/interseismic interval of a few decades. In addition a limited number of geodetic observations provide information about the deformation that was occurring before the earthquake. This article is a critical review of these surface observations and of the numerical models that have been employed to explain them.

Most, although not all, of the deformations that are relevant to the earthquake cycle have been observed through the use of geodetic survey techniques. Tide gauge measurements, along with other geodetic observations also provide important data. Thus, our primary focus will be on the time scales associated with these observations, typically a few months to several decades. Also essential is the much shorter time scale associated with the earthquake itself and the somewhat longer time scale associated with the recurrence of great earthquakes. Although, we will be less concerned with the very long term development of geological features these features are the product of both seismotectonic and geomorphological processes. Thus the insights provided by geological investigations supplement what can be learned from geodetic and seismological investigations.

Since we are discussing the 1964 PWS earthquake, it is appropriate to begin this review with a few basic facts about the earthquake itself. The earthquake was, and remains, the second largest seismic event to have occurred during the modern instrumental era. Its moment magnitude (M_w =

9.2) is exceeded only by that of the 1960 Chile earthquake ($M_w = 9.5$). The rupture extended over a zone that was 600-800 km long and 200-250 km wide (**Fig. 1**). The shaking produced by the earthquake was felt at some locations for 4 to 5 minutes. The rupture began just north of Prince William Sound; the epicenter was $61.04 \pm 0.05^\circ\text{N}$ and $147.73 \pm 0.07^\circ\text{W}$ (Sherburne *et al.*, 1972). The motion was primarily a thrust on a low dip-angle plane but there may have a strike-slip component to the motion. As there were no seismometers located close to the epicenter and the initial seismic signals saturated many of the high gain global-network seismographic instruments then in operation, the determination of focal depth was challenging (Stauder, 1972); however, the focal depth estimate of 25 ± 10 km by Oleskevich *et al.* (1999) is probably good. The best constraints on the source depth may come from a combination of the horizontal position of the aftershock zone and depth profile of the downgoing plate. This gives a depth of 20-25 km. Certainly, the depth of the down-dip limit of the coseismic rupture surface appears to have been not more than about 40 km, in agreement with the depth for a magnitude 6.9 aftershock that occurred in 1965 (Tichelaar and Ruff, 1993). This limit is shallower than the 350°C - 450°C temperature contour computed and used by Oleskevich *et al.* (1999) as the down-dip limit for great earthquakes. The up-dip depth limit is about 4-6 km, and corresponds to the 100°C to 150°C temperature contour suggested by the same authors as an appropriate up-dip temperature limit. The rupture lasted about 100 seconds in the Prince William Sound area and about 40 seconds at Kodiak Island (Christiansen and Beck, 1964). For the first 44 seconds of the event, the rupture propagated in different directions, but after that it propagated primarily to the southwest (Wyss and Brune, 1972). The region of the eastern Aleutian-Alaska arc that ruptured in 1964 had not experienced a great earthquake within the historical instrumental period dating back about 65 years, although four earthquakes of magnitude 7.8-8.3 occurred near Yakutat Bay, to the east and

in a different tectonic environment, during the 13 month period from September, 1899 to October, 1900. In addition five earthquakes of magnitude 7.0-7.2 occurred between 1928 and 1957 within the 1964 rupture zone. We will have more to say about the paleoseismic record of great earthquakes in this region later in this article.

As a geodetic event, the earthquake produced static coseismic displacements estimated to have exceeded 20 m horizontally in some locales and reaching about 10 m vertically at other locations (Toucher, 1972; Plafker, 1972; Parkin, 1972). Most of the displacements were associated with slip on the plate boundary megathrust although some of the larger vertical motions are thought to be associated with local slip on subsidiary faults (Plafker, 1972). The megathrust itself has considerable complexity, Note only does the dip angle vary along strike, but the nature of the boundary itself also changes. In the western and central portions of the rupture zone, the megathrust is the boundary between the North America and Pacific Plates while in the Prince William Sound region it is the interface between the North America Plate and the Yakutat Block, an allochthonous terrane incompletely welded to the top of the Pacific Plate.

Among the geodetic techniques that have been used to measure crustal deformation in southcentral Alaska are triangulation, leveling, and other conventional survey techniques and space-age measurement systems such as Very-Long-Baseline Interferometry (VLBI) and the Global Positioning System (GPS). As later discussions will illustrate, GPS observations are providing an especially detailed picture of the contemporary three dimensional surface crustal movement. This picture reveals aspects of the crustal deformation cycle that have not been appreciated until recently but may be common in Alaska and other subduction zones. While GPS measurements are providing much of today's data, other measurements have value from both a contemporary and historical perspective. For example, tide gauge observations of apparent sea

level change can be used to study the crustal movements in those locales where the change in sea level is due primarily to the vertical movement of the land on which the instrument is located rather than a change in the height of the ocean. In many cases tide gauge measurements give the longest available record of crustal movement. Specifically, at Kodiak city and Seward tide observations began before the earthquake and provide a fairly continuous record of preseismic and postseismic vertical motion. At other, temporary tide gauge sites, measurements were made before the earthquake and again shortly afterwards and provide information on the coseismic vertical motion. Several more permanent tide gauge stations began recording sea heights shortly after the earthquake.

The geodetic measurements of crustal movement span a time interval that is only a fraction of the typical recurrence time for a great earthquake. Additional information about crustal movements come from geologic studies that estimate such long-term parameters as slip rates on faults and uplift rates for geologic features, as well as surface displacements in specific earthquakes. In the case of the PWS earthquake, the mapping of barnacle and alga lines provided much of the information about the coseismic vertical motion while alternating deposits of mud and peat have provided information on both interseismic sedimentation and coseismic uplift or submergence (Carver and McCalpin, 1996). Paleoseismic information is also invaluable for developing a chronology of past earthquakes (Bartsch-Winkler and Schmoll, 1992).

There are at least two compelling scientific reasons for undertaking the types of crustal deformation studies reviewed in this article. The first is improve the understanding of how the Earth prepares for an earthquake and relieves the stress accumulated by the resistance to plate motion at plate boundaries. The second is to infer geophysical parameters describing the interior of the Earth that aid in the understanding of a variety of geophysical processes both related to and

distinct from the earthquake process itself. Concerning the first of these two objectives, there are two major paradigms that provide the conceptual framework for contemporary crustal deformation studies. The older of these is a deterministic view based on the idea that earthquakes can be understood by considering the slow accumulation and sudden release of stress at a fault or within a fault system. The second, more recent paradigm, is that earthquakes are better understood by considering the statistical properties of a nonlinear chaotic system in a near critical state (Rundle *et al.*, 2000). The spatial scale involved in earthquake studies ranges from the microscope scale (10^{-6} to 10^1 m) associated with rock frictional properties to the tectonic plate boundary scale (10^5 to 10^7 m). To some extent the paradigm most useful for a specific study depends on the spatial and temporal scale of interest, but the deterministic paradigm is the philosophic basis for most of the studies reviewed here. The most elementary form of this paradigm is the elastic rebound theory in which both stress and strain accumulate at a constant rate between earthquake and are relieved by the sliding associated with the sudden overcoming of frictional resistance to motion. As we shall see, this simple version of the paradigm is largely being rewritten by studies such as those in Alaska that reveal that temporal variability and spatial heterogeneity in the deformation process is an integral part of the earthquake cycle. The complex temporal and spatial characteristics of the southern Alaska subduction zone are manifest in the occurrence of secular and transient strain accumulation, quasi-static slip processes, a variety of relaxation times for postseismic rebound, and various strengths of coupling between the overthrust and subducting plates.

Comprehensive study of the PWS earthquake began shortly after the earthquake. Many of the early studies were later compiled into an eight volume report, "The Great Alaska Earthquake of 1964," that was published by the National Academy of Sciences in 1972. The report has distinct

volumes devoted to Geology, Seismology and Geodesy, Hydrology, Biology, Oceanography and Coastal Engineering, Engineering, Human Ecology, and Summary and Recommendations. We note that several of the papers cited in this article can be found in the second of these volumes, although in some cases the papers are copies or updates of earlier publications. As an examination of the extensive reference list of this article will confirm, studies of the crustal deformation associated with the earthquake cycle in southcentral Alaska have continued to engage the attention of the scientific community in the decades since the earthquake.

2. Tectonic, Geologic, and Seismologic Setting

The contemporary tectonic environment of southern Alaska is shown in **Fig. 2**. At locales west of Prince William Sound, the Pacific Plate subducts under the North America Plate with a small to moderate degree of obliquity. At 57.5°N, 148°W, a locale near the eastern end of the Alaska/Aleutian Trench, the velocity of the Pacific Plate relative to North America, as estimated from geological data, is about 55 mm/yr at N19°W (DeMets *et al.*, 1990, DeMets *et al.*, 1994). Newer estimates of Pacific-North America relative plate motion based on contemporary geodetic observations give a similar velocity (e.g. DeMets and Dixon, 1999; Sella *et al.*, 2002). The velocity increases slowly, and rotates westward as one progresses to the southwest and is about 57 mm/yr near 56°N, 152°W. The strike of the Alaska-Aleutian trench varies with location, but gradually rotates clockwise from its southwest-northeast orientation near its eastern terminus, southeast of Cordova (Beckman, 1980) to an east-west orientation near the middle Aleutian island chain. The obliquity of the plate motion vector relative to the trench normal is about 27° at the Kenai Peninsula, but is less than half that at Kodiak Island. Thus, the relative plate motion in

southcentral Alaska is approximately oriented normal to the trench and there is general agreement that the normal component of the plate motion is accommodated by slip on the megathrust and splay faults. There is, however, some debate about the extent to which motion parallel to the plate boundary is partitioned between slip on the megathrust and strike-slip crustal faults. If slip partitioning were active, left-lateral strike-slip faulting should occur in the North America plate. However, most crustal strike-slip faults in Alaska display right-lateral motion. Thus Williams *et al.* (2001), for example, concluded that slip partitioning was not active in this region. The 1964 earthquake may have involved some left-lateral slip, as well as the dip-slip motion, on both the megathrust and offshore faults. This issue will be discussed in more detail when we summarize observations and models of the coseismic motion.

The geometry of the plate interface varies along the strike of the plate boundary. In the PWS region, the plate interface dips at a shallow angle, the dip being 3° to 4° (Brocher *et al.*, 1994; Doser *et al.*, 1999). As already mentioned, in the PWS region the plate interface is between the North America Plate and the Yakutat Block, an allochthonous continental terrane lying over the Pacific plate and partly welded to it (Perez and Jacob, 1980; Plafker *et al.*, 1994b). Underneath the Yakutat Block the top of the Pacific Plate dips at an angle of 8° - 10° to the horizontal. Similarly the dip angle of the locked interface between the Pacific Plate and North America is about 8° near Kodiak Island (von Huene *et al.*, 1987; Fletcher *et al.*, 2001), a locale that is hundreds of kilometers west of the western edge of the Yakutat Block.

As for crustal faults, the Denali fault is a right lateral fault that extends for 500 km from near the Fairweather fault system southeast of the rupture zone of the PWS earthquake to under Denali Mountain (Mt. McKinley), well to the north of the 1964 rupture zone. Slip on this fault has been ~ 10 mm/yr in geologically recent times (Plafker *et al.*, 1994b), and GPS results show a

similar rate of 7 – 11 mm/yr for the present motion (Fletcher and Freymueller, 2002). Line length measurements from the 1980's showed a slower rate (Savage and Lisowski, 1991), but these earlier results are not entirely inconsistent with the GPS and geological estimates given the large uncertainties in the line length data. In 2002, a magnitude 7.9 right-lateral strike slip earthquake occurred on a portion of the Denali-Totschunda fault system located south of Fairbanks. This is the largest documented strike-slip event on the fault system and clearly demonstrates that this fault system is actively responding the contemporary tectonic motion, although the details are not known. There are several other right lateral faults in southern Alaska north of the subduction zone that accommodate block rotation and shearing within interior Alaska. However their relationship to contemporary tectonic processes including the plate boundary subduction processes, the collision of North America with the Yakutat block, and southeast and northern Alaska tectonics is not yet understood in full detail (Yeats *et al.*, 1996).

The geometry of the plate interface is an important parameter for describing the subduction zone structure. Zweck *et al.* (2002a) derived a detailed description of the plate interface in southcentral Alaska by applying a two-dimensional cubic spline interpolation (with a resolution of 5 km) to the combined plate interface depth information of Doser *et al.* (1999), Page *et al.* (1991), Wolf *et al.* (1991), and Moore *et al.* (1991). They used the digital elevation/bathymetry model, ETOPO5, to estimate the interface depth near the trench. The interface depth contours emerging from their analysis are shown in **Fig. 3**. Notice the considerable deflection of the contours on, and north of, the Kenai Peninsula and the much stronger gradient in the mid-depth contours near Kodiak Island than on the Kenai Peninsula. The 20 km contour passes near Prince William Sound, through portions of the eastern and central Kenai Peninsula and is near the trenchward coastline of Kodiak Island. The steepness of the plate interface increases rapidly once

the slab surface reaches 40-50 km depth (Ponko and Peacock, 1995). Another important parameter of the subduction zone is the subducting lithospheric plate thickness. Zhao *et al.* (1995) estimated that the Pacific Plate thickness near the plate boundary to be 45-55 km.

The geology and tectonic framework of the Alaskan plate boundary region and the interior is comprehensively discussed in the compendium publication, "The Geology of Alaska," published by the Geological Society of America. Here we summarize some salient features of the southcentral Alaska region, as presented primarily in the chapter by Plafker *et al.* (1994b). The Pacific plate, (and before that the Kula plate) has been moving northwest relative to North America during the past 55 million years. The motion in southcentral Alaska has been primarily accommodated by underthrusting along the Aleutian Arc and by right lateral to oblique-right lateral slip on the Queen Charlotte-Fairweather Fault system to the east. Some motion has also been taken up by slip on the bounding faults of the Yakutat terrane within Alaska. The crust of Pacific Plate is progressively older going from east to west along the plate boundary. It is about 37 Ma at the eastern end of Aleutian trench and 45 Ma near Kodiak Island (Atwater and Severinghouse, 1989). The continental crust is made up of a variety of terranes that have been accreted onto the North America margin. A major structural boundary of the region is the Border Ranges Fault System (BRFS) which is a suture between the Chugach terrane lying to the south and the Peninsular terrane to the north (**Fig. 2**). The terranes north of the BRFS have island arc and oceanic plateau origins (Pavlis and Crouse, 1989 and references therein) while those to the south are from a subduction-related accretionary complex. The BRFS played a role in accommodating the subduction during the early Jurassic and late Cretaceous and may have later been a site of lateral transport (Oldow *et al.*, 1989), although this latter point is disputed (Little and Naeser, 1989). The northern portion of the fault bent in response to the counterclockwise

rotation of western Alaska in the Eocene. While the fault is a major structural boundary, evidenced, for example, by a sudden change from a mountainous to flat environment on the Kenai Peninsula, it exhibits little or no contemporary seismic activity. No fully conclusive evidence for Holocene displacement been found (Wolf and Davies, 1986), although there is some limited evidence for motion near Anchorage within the past few hundred years (Updike and Schmoll, 1985). The major features of the Chugach terrane are the Chugach and Kenai Mountains and most of Kodiak Island. South of the Chugach terrane is the Prince William terrane. Another major structural feature in southcentral Alaska is the Contact Fault System (CFS). The CFS marks much of the boundary between the Chugach terrane and Prince William terrane to the south. On the eastern side of the region considered in this article, the Yakutat terrane has been displaced about 600 km northwest along the Queen Charlotte-Fairweather Fault System (Plafker *et al.*, 1994b).

Most of the interesting physical features shown in **Fig. 2** are discussed by von Huene *et al.* (1999). They include the aforementioned crustal faults and terranes and several subducting seamounts, some of which act as asperities. The major asperities are reflected in the coseismic slip pattern and in the location of aftershocks. However, some of the details on the asperity geometry are not well constrained or universally accepted. The subduction zone in the Kodiak region, unlike Prince William Sound, has a large accretionary prism. **Fig. 2** also shows the location of three seismic reflection and refraction lines (TACT, EDGE, and ALBATROSS) from which much of the information about the structure of the southcentral Alaska subduction zone has been derived (e.g., Moore *et al.*, 1991).

As to the contemporary seismicity of southcentral Alaska, the locations of large earthquakes with focal depths less than 60 km in southcentral Alaska occurring since 1899 (as obtained from

the catalog of the Alaska Earthquake Information Center) are shown in **Fig. 4**. We note, in particular, a number of events near Kodiak Island. There is a scarcity of major earthquakes between the Kenai Peninsula and the Afognak and Kodiak Islands. Both Page *et al.* (1991) and Doser *et al.* (1999) have noted that since the PWS earthquake there have been few earthquakes on that portion of the plate interface that slipped during the event. They found three clusters of moderate-size earthquakes that were associated with the subduction zone, but did not occur on the plate interface. The first two clusters were in the vicinity of the Prince William Sound asperity with one involving reverse or low angle thrusts and the other, normal to oblique faulting within the subducted Pacific Plate. The third cluster of events was in the Cook Inlet at depths below the plate interface. There is wide-spread seismicity within the overriding continental plate. In particular, Page *et al.* (1991) noted a band of seismicity (earthquakes with magnitude 3.0 or greater and depths less than 40 km occurring from 1967 to June, 1988) north of and paralleling the Castle Mountain fault east of 151°W and another seismic belt along the volcanic axis west of Cook Inlet. They also noted a diffuse north-northeast trending band of earthquakes near 150°W longitude between Cook Inlet and the Denali Fault. The shallow seismicity on the Kenai Peninsula does not correlate well with mapped fault traces, but shocks tend to concentrate beneath the Kenai mountains and are, therefore, east of the Border Ranges Fault (Stephens *et al.*, 1990).

As we mentioned above, some of the seismic asperities in the southcentral Alaska subduction zone may be associated with subducting seamounts. Estabrook *et al.* (1994) have noted that the northeast limit of five magnitude 7 earthquakes that occurred in 1938 appears to coincide with the southwest limit of the 1964 rupture. They suggest that this segmentation boundary is associated with a warping of the Pacific Plate and a change in dip angle from around 8° near

Kodiak Island to about 19° to the southwest. However, the evidence for such a dramatic change in dip angle is not totally compelling. Fletcher *et al.* (2001), using the reflection profiles of von Huene *et al.* (1987) and Moore *et al.* (1991) and source models of the 1938 Mw 8.3 earthquake, estimated a subducting plate dip angle of 6° for the region southwest of Kodiak Island. Von Huene *et al.* (1999) suggested that the differences between the Prince William and Kodiak asperities is related to differences in the surface properties of the lower (i.e. subducting) plate in the two regions. These authors argued for a relationship between the Kodiak asperity and the subducted Kodiak-Bowie hot spot swell. They also related the region of low aftershock energy release between the Kodiak and Prince William asperities to the thickened trench sediment adjacent to the trailing edge of the Yakutat Terrane. The region around Prince William Sound, near the 1964 epicenter, is particularly complex from a seismotectonic perspective with the direction of maximum horizontal compression showing considerable variability (Estabrook and Jacob, 1991).

3. Observed Crustal Motion

Although there is a certain logic to discussing the observed crustal motion in temporal sequence by considering the preseismic motion followed by the coseismic, postseismic, and interseismic motion, we choose the alternative of discussing the coseismic motion first. This approach is pedagogically useful for an awareness of the deformations that accompanied the earthquake provides a useful conceptual basis for understanding the motions that preceded and followed the event.

3.1 Coseismic Crustal Motion

The coseismic motion associated with the 1964 earthquake has been discussed by several

workers. As far as geodetic observations are concerned, both triangulation data (Parkin, 1972) and leveling data (Small and Wharton, 1972) are available. The coseismic horizontal displacements were obtained by differencing post-earthquake observations made in 1964 and 1965 with pre-earthquake observations made between the early 1900's and the late 1940's. There is considerable spatial variation in the quality of the pre-earthquake survey data. The region from Anchorage to Glennallen and then to Valdez was surveyed to first-order standards. An arc across northern Prince William Sound to Perry Island (south of Valdez) was surveyed to second-order standards. An arc from Perry Island to Anchorage was surveyed to third-order standards. A portion of the Kenai Peninsula, including an arc from Seward to Turnagain Arm, was also surveyed to third-order standards. For the most part, the observations were made for chart control rather than crustal movement studies. There are substantial misclosures for some of the triangulation arcs. Because the pre-earthquake data span four decades, part of the misclosure error is due to deformations that occurred before the earthquake. Data from the third order arc along Turnagain Arm are particular suspect (Parkin, 1972). The scale for the pre-earthquake network was provided by a limited number of taped distance observations and a single microwave (tellurometer) distance observation. Orientation was provided by several Laplace azimuth observations. The quality of the post-earthquake observations is more uniform than that of the pre-seismic surveys with the observations being a mixture of first order triangulation and tellurometer observations. Parkin (1972) made a free adjustment of the network to determine the magnitude and direction of coseismic slip vectors. He used 1476 observed directions at 292 stations, 5 Laplace azimuths for orientation, and 8 geodimeter and 146 tellurometer observations for scale. The errors in the station positions estimated from an examination of residuals are 2-3 m in the Prince William Sound region. The results that were presented by Parkin (1972) are relative

to a fixed site (Fishhook) north of Palmer Alaska, at $61^{\circ} 43.0' \text{ N}$ and $149^{\circ} 14.0' \text{ W}$. These adjusted coseismic displacements are shown in **Fig. 5**. The maximum observed displacements were in excess of 20 m and were directed south-southeastward. As expected, displacements were generally largest near the coast and less at locales further inland. There were some exceptions to the systematic variations in displacement vectors with distance from the trench. For example, the displacement at Middletown Island, the most southerly locale at which observations were made, did not follow the general pattern. The displacement here most likely reflected both reduced coseismic slip along the megathrust at very shallow depths and the influence of slip on the subsidiary Patton Bay fault lying in the crustal over-thrust block. Furthermore, the absolute motion of the entire network had some ambiguity as the displacements were calculated relative to Fishhook (**Fig. 5**), a site that may not have been motionless. As Whitten (1972) mentions, "In Parkin's investigation the western end of this scheme was held fixed....The eastern end was permitted to seek its most probable position." While it appears that the direction of many of the coseismic slip vectors were aligned with the relative plate motion vector and that there was an easterly component to the motion at sites on the eastern side of the network, this may be an artifact of the analysis. If, alternatively, the displacement of these easterly sites was in the direction of plate motion, then there was a left lateral component to the largest of the displacement vectors. It is not known whether the northwesterly coseismic motion of sites on the western side of the Kenai Peninsula was real or whether it was an artifact of a weak network linkage to this region. Curiously, there is a dramatic difference in the direction of interseismic motions of the eastern and western portions of the Kenai Peninsula as well. We will discuss this difference in the direction of interseismic motion in detail later.

An illustrative example of the geometry the triangulation networks, specifically that of the

southcentral portion of the triangulation network is shown in **Fig. 6**. Data from this and other portions of the triangulation network were used by Pope (1972) to compute strains. The network was less than ideal in many places for it contained both long linear segments and triangles with large inter-site separations; however, the general results from the analysis were reliable. The direction of the principal coseismic strain axis was north-northwest in the Seward region, but north-northeast in the tectonically complex region near Prince William Sound. We do not show the diagrams of the strain field here because the published figures do not reproduce well, but the originals can be found in Pope (1972).

As for the coseismic vertical motion, five repeated leveling profiles and sea level heights measured by tide gauges at Homer, Seward, Whittier, and Valdez were used to define the surface displacements (Small and Wharton, 1972). Among the leveling profiles, the most interesting is one that ran from Matanuska through Anchorage, then southward through Portage along the Turnagain Arm and extended from there southward to Seward along the Alaska railroad right-of-way. The pre-earthquake survey was conducted in 1922 and the post-earthquake survey in 1964. Coseismic subsidence was noted along the entire profile, with a vertical height change of ~ 1 m at Seward and a maximum subsidence of ~ 2 m south of Portage (**Fig.7**) Another resurveyed line ran west-northwest from Matanuska through Glennallen then turned to the north. The pre-earthquake survey was conducted in 1944. Along this route the typical coseismic subsidence was about ~ 0.5 m and there was only a modest variation in the subsidence along the route. This small variation in uplift is not surprising since the survey line ran roughly parallel to the trench. Along a profile running initially southward from Glennallen and then turning eastward toward Valdez, first surveyed in 1923, the maximum coseismic subsidence exceeded 1.2 m. The measured subsidence was less than 1 m at Valdez but there was a 2-3 m difference in the tidal datum as

determined in May through July of 1964 versus July through November. The difference is too large to be attributable to postseismic motion.

Plafker (1972) developed a contour map of the vertical coseismic displacements throughout southcentral Alaska by combining the leveling data, tide gauge observations, the changes in the upper growth level of intertidal organisms, changes in the heights of storm beaches, changes in the tidal benchmark heights relative to sea level, and estimates of changes in the location of the shoreline by local residents. His generalized map is shown in **Fig. 8**. Several features should be noted. The contours are oriented southwest to northeast, a direction consistent with the strike of the trench, the trend of the surface geological features, and the orientation of magnetic and gravitational anomalies. The oceanward portion of the zero uplift contour passes near the southern edge of Kodiak Island, south of most of the Kenai Peninsula and near to Valdez. Cordova, Middleton Island, and much of the Gulf of Alaska is in the region of coseismic uplift. The maximum coseismic uplift was around 10 m (n.b., the Plafker map units are feet). The uplift contours are influenced by localized slip on the Patton Bay and other shallow reverse faults. Most of Kodiak and Afognak Islands, the Kenai Peninsula, the Matanuska Valley and the region north of Prince William Sound was in the zone of coseismic subsidence. The width of the subsidence region broadened considerably around Anchorage. The maximum coseismic subsidence was around 2 m. Plafker (1972) suggested that a region of modest uplift occurred inland beyond the zone of subsidence, notably in the Aleutian range northwest of Cook Inlet and north of the Talkeetna Mountains between Anchorage and Fairbanks.

One of the intriguing aspects of the coseismic vertical deformation pattern comes from terrace data at Middleton Island. The observed 3.3 m uplift (Parker, 1972) is apparently only about half that for the preceding three earthquakes (Plafker, 1972). There has been speculation that

additional uplift will occur in a future rupture of the eastern segment of the 1964 earthquake zone or that there was a greater partitioning of slip onto the crustal faults in the 1964 earthquake than in preceding events (Carver and McCalpin, 1996).

Information on the coseismic and horizontal surface coseismic displacements have been used to develop models of fault slip during the earthquake. Hastie and Savage (1970) performed a least-squares analysis using an elastic dislocation model with a single fault plane on the plate interface and a subsidiary slip surface on Montague Island. They estimated the slip on the main thrust to be 10 m. Their work, along with that of many others (e.g. Plafker, 1965; Stauder and Bollinger, 1966; Kanamori, 1970), helped establish that the main fault plane was a low-dip angle megathrust rather than the nearly vertical fault extending to 100-200 km that had been proposed earlier by Press (1965) and Press and Jackson (1965). Although this latter view, i.e., that the earthquake slip might have been primarily on a vertical fault extending to great depth, might seem strange to the present-day reader, it was not an easily dismissible hypothesis in 1964 when plate tectonics was not a well established paradigm (see, e.g., Harding and Algermissen, 1972).

Miyashita and Matsu'ura (1978) developed a more complex coseismic deformation model in which they adopted constraints from the vertical displacement data. Their major contribution was to consider the slip distribution over multiple (four) surfaces rather than a single plane. Three of their slip planes were along the main thrust, laterally adjacent to one, and the fourth was at Montague Island. The three main thrust fault planes were centered on Kayak Island (on the eastern side of the rupture), Prince William Sound, and the Kenai Peninsula-Kodiak Island region. Miyashita and Matsu'ura (1978) inverted the model to estimate slip, dip angle, strike, and fault segment extent. The Kodiak-Kenai segment had the largest estimated slip, nearly 19 m, while the Montague Island segment had an estimated slip of 11.5 m. More recently, Holdahl and

Sauber (1994) developed a detailed coseismic slip model based on a variety of previously published data including (1) changes in coastal heights from barnacle, vegetation, storm beach, and marker observations, (2) change in tidal datum planes, (3) pre- and postseismic leveling, (4) pre- and postseismic bathymetry, and (5) horizontal displacements. Their inversion model allowed for both strike-slip and dip-slip components of the motion, but the inferred strike-slip displacements were generally small except in segments near Prince William Sound and northeast of there. There was an area of high displacement near the coast and offshore from the Prince William Sound region through the eastern Kenai Peninsula. The highest slip was about 30 m, near Montague Island. Johnson *et al.* (1996, see also Johnson, 1999) updated and improved on the Holdahl and Sauber (1994) analysis, conducting a joint inversion using 23 tsunami waveforms as well as 188 vertical and 292 horizontal geodetic observations in their inversion. Their model had only 18 fault segments, but had the meritorious feature that the dip angles and depths for segments near the Yakutat terrane were consistent with locating the megathrust at the top of that terrane rather than the top of the Pacific plate. They adopted Holdahl and Sauber's (1994) estimate of 8.5 m slip for the Patton Bay fault at Montague Island. Despite some differences in detail, the slip distribution deduced by Johnson *et al.* (1996), shown in **Fig. 9** closely resembles that of Holdahl and Sauber (1994). The peak fault displacement was 22.1 ± 4.4 m. One of the most important aspects of the Johnson *et al.* (1996) analysis was full consideration of tsunami data. A significant portion of the interface slip occurred offshore on southcentral Alaska, removed from locations where geodetic measurements could be made. Thus, geodetic survey data alone do not provide sufficient constraints on the slip pattern, particularly at locations near the trench. Holdahl and Sauber (1994) used tsunami information as *a-priori* inputs to their model, but Johnson *et al.* (1996) used the full strength of the tsunami data.

Christensen and Beck (1994), and earlier Kikuchi and Fukao (1987) studied the seismic moment release of the 1964 earthquake. Both pairs of authors identified an area of large moment release near Prince William Sound, although not centered on the epicenter. In addition, Christensen and Beck (1994) identified a secondary region of moment release near Kodiak Island. They found only modest moment release between the two locales (**Fig. 10**). Similarly, Johnson *et al.* (1996) emphasized that their inversion analysis produces indicates large and secondary amounts of slip at the Prince William Sound and Kodiak asperities, respectively. The comparatively small moment release between the Prince William Sound and Kodiak asperities may be related to a tear in the subducting Pacific Plate. Pulpan and Frohloch (1985) argued that such a tear exists underneath Cook Inlet at about 59°N.

One issue that remains a matter of some debate is the magnitude and location of any strike-slip motion that might have occurred during the PWS earthquake. Hastie and Savage (1970) considered slip on both the main thrust and a subsidiary fault running through Montague Island. They estimated a left lateral slip of nearly 10 m on the main thrust and 12 m on the subsidiary fault. Miyashita and Matsu'ura (1978) also found significant left lateral motion on the subsidiary fault but relatively little lateral motion on any of the three faults that characterize their main thrust. Holdahl and Sauber (1994) found right lateral slip on the main thrust in the vicinity of Prince William Sound, but this result appears anomalous. Johnson *et al.* (1996) constrained the slip to be in the dip direction along much of the plate interface and in the direction of plate motion (giving a left lateral component to the slip) in the vicinity of Prince William Sound.

The recurrence time between major earthquakes is a key parameter for assessing earthquake hazards as well as modeling the temporal evolution of the crustal deformation. Based on an average coseismic displacement of 14 m and a somewhat high North America-Pacific relative

plate velocity of 65 mm/yr, Nishenko and Jacob (1990) estimated a recurrence interval of 233 yrs for the Prince William Sound segment of the plate boundary. Using a peak displacement of 30 m, rather than the average displacement, they derived an alternative recurrence period of 462 yrs. This second estimate is closer to the geological estimate, as we will summarize below, but might be low if some of the convergent motion between North America and the Pacific is accommodated aseismically. As to geologic estimates of recurrence rates and the time since the last earthquake, Nishenko and Jacob (1990) quoted radiocarbon dates of observed terraces, notably at Middleton Island indicating a recurrence interval of 1021 ± 362 yrs. The most recent terrace at Middleton Island dates from 1300 years ago. However, the Middleton Island estimate may not be representative. Combellick (1993) cited data obtained along Turnagain and Knik Arms of Cook Inlet giving an average recurrence interval of 590-780 yrs which would be in closer agreement with the longer of the estimates from geodetic data. Similarly, Plafker *et al.* (1992) reported an age of 665-895 yrs for the most recent earthquake uplift in the Copper River delta. Combellick (1992) obtained evidence from Girdwood, along the Turnagain Arm that the penultimate great earthquake in the Anchorage region occurred 700-900 years ago and there is evidence from peat stratigraphy in a borehole near Portage that some significant earthquakes may have occurred within a century of one another (Combellick, 1991).

For the Kodiak Island segment of the plate boundary, Nishenko and Jacob (1990) derived a recurrence interval of only 50-60 yrs, but the magnitude of Kodiak Island earthquakes was in the range $M_s = 7.5 - 8$, while the PWS earthquake had a much larger magnitude, so the different recurrence intervals also refer to different earthquake magnitudes as well.

3.2 Preseismic and Interseismic (pre-1964) Crustal Motion

Tide gauges located at Seward and Kodiak city give some limited insight into the vertical

motion in southcentral Alaska in the decades preceding the 1964 earthquake. The tide gauge at Seward dates back to 1925 and the one near Kodiak city to 1950. Savage and Plafker (1991) examined the pre-earthquake tide gauge records at these two locales and deduced that the mean uplift rate before the earthquake was 2.7 ± 1.5 mm/yr at Seward and 4.8 ± 1.6 mm/yr at Kodiak. These modest uplift rates are consistent with Plafker's (1972) claim that "no change in mean sea level attributable to tectonic movement was detected in a 21 year tidal record at Seward and a 15 year record at Kodiak." The expression, "no change" must be taken in proper context. The statement was written shortly after its author had measured vertical motions associated with the earthquake that were orders of magnitude larger than any known preseismic motion; thus any preseismic motion was not considered to be of consequence. However, Shennan *et al.* (1999) have documented evidence for preseismic subsidence by examining microfossil assemblages along Turnagain Arm. As to postseismic motion, the uplift rate at Kodiak after the earthquake was much larger than the preseismic rate (as discussed in more detail below) suggesting either that there is a time dependence to the uplift rate over the earthquake cycle or that the deformation rate during one interseismic period is very different than in another.

Plafker (1972) also discussed the recent Holocene record of shoreline movements. A large number of dated shoreline samples have radio-carbon dates of less than a 1000 years before the 1964 earthquake and thus may reflect preseismic and interseismic deformation. Of these, all of the sites that are in the zone of coseismic uplift appear to be located below the pre-1964 sea level. This suggests that these shoreline points gradually submerged during the interseismic period. However, samples with older carbon dates reflect a pattern similar to the coseismic deformation. Thus, several points on the Kenai Peninsula and Turnagain Arm that subsided during the earthquake show Holocene subsidence as well. There is strong evidence for

emergence of sites from Middleton Island and near Bearing Lake between Cordova and Yakataga, locales that rose during the earthquake. It appears that the geologically recent vertical motion due to tectonic processes is more significant than that due to postglacial rebound at most locales, but the net emergence or submergence is complex combination of the coseismic and interseismic motions. Plafker (1972) states that this interplay has resulted in the net emergence of part of the continental margin and simultaneous submergence of part of the Kenai-Kodiak mountains.

3.3 Postseismic and Interseismic (pre-1964) Crustal Motion

An early report of significant postseismic motion based on geodetic observations was published by Small and Wharton (1972). They compared leveling observations along Turnagain Arm made in October 1964 to those made earlier, in May and June of that year. As shown in **Fig. 11** the observed motion was a trenchward tilting. Small and Wharton (1972) interpreted the tilting as being an indicator of subsidence at Portage, using as additional evidence the results of a survey taken in 1965 which they felt demonstrated further subsidence at Anchorage and Portage. However, as the later results discussed in Brown *et al.* (1977) demonstrate, the Turnagain Arm observations are more convincingly interpreted as an indication of significant postseismic uplift along much of the route between Portage and Anchorage. Cohen (1998) showed that the uplift deduced from the two surveys made six months apart in 1964 could be extrapolated to predict the uplifts observed over a year interval as derived from the initial survey and one made in 1965. This result indicates that the rapid, early postseismic uplift proceeded at a roughly constant rate for at least one year following the earthquake. The maximum relative uplift rate along Turnagain Arm in this first year was about 150-160 mm/yr, although it is possible that the absolute uplift rate was somewhat less. Brown *et al.* (1977) examined both leveling and tide gauge observations

for indications of postseismic vertical motion. The leveling data was from surveys made along Turnagain Arm in 1964, 1965, 1968, and 1975. The observations (**Fig. 12**) indicated a maximum uplift between 1964 and 1975 of 37 cm relative to the tide gauge station at Anchorage. The maximum uplift occurred at D73, a site located approximately 40 km southeast of Anchorage along the 100-km long Turnagain Arm profile. The relative uplift dropped to about zero near Portage, located 75 km southeast of Anchorage. Modest relative subsidence occurred between there and Whittier. To estimate absolute rather than relative motion, Brown *et al.* (1977) used tide gauge observations of the uplift rate of Anchorage. They deduced that the Anchorage tide gauge site was moving upward at a constant velocity of 16.3 mm/yr in the years following the earthquake. If this were correct, then the absolute uplift of D73 by 1975 was in excess of 50 cm. However, by analyzing a longer time series at the Anchorage tide gauge, Savage and Plafker (1991) concluded that there has been little net vertical motion at Anchorage since the earthquake. The uplift rate they derived was a slow -0.7 ± 1.3 mm/yr (post-earthquake, but prior to 1973) or 1.0 ± 2.2 mm/yr (1974 to 1988). Applying a constraint of no postseismic uplift at Anchorage to the leveling data means that the aforementioned relative uplifts are also absolute uplifts, i.e., the maximum absolute uplift (through 1975) was about 37 cm. The analysis of the Anchorage tide gauge data also revealed that there may have been significant oscillatory transients in the years immediately following the earthquake, an observation we will return to briefly below.

Another intriguing aspect of the leveling observations is the time dependence of the uplift (**Fig. 13**). Again taking Anchorage as a reference point, the maximum uplift rate for the first year after the earthquake was 80-90 mm/yr (150-160 mm/yr relative to Portage). For the four years between 1964 and 1968 the maximum uplift rate decreased to about 50 mm/yr, and between 1964 and 1975 it was about 30 mm/yr. However the location of the point of maximum uplift appeared to

migrate somewhat to the southwest, i.e. trenchward, during this period. Perhaps more revealing is the time history of individual pair of sites. For example, the uplift rate of site D73 relative to Anchorage decayed with an exponential relaxation time of 5.2 yrs. Other sites also manifest relaxation times of a few years. The observation that the transient decayed over a few years is significant because data that we will discuss below reveal longer-lived postseismic transients. It is likely that different relaxation times are reflective of different mechanisms or at least that the parameters governing the relaxation vary with locality.

Prescott and Lisowski (1977, 1980) examined a set of leveling observations on Middleton Island, located about 60 km northwest of the Alaska-Aleutian Trench. The observations were made in 1966, 1974, and 1975. They indicate a down-to-the northwest tilt. The tilt rate based on the first two surveys was $11.5 \pm 5.3 \mu\text{rad/yr}$, but the rate based on the last two surveys was $5 \pm 0.8 \mu\text{rad/yr}$. Thus, there is a suggestion that the tilt rate may have decreased with time, but the details of the temporal evolution are not known. A critical aspect of the interpretation of data from Middleton Island is that the deformation there may be strongly influenced by strain accumulation and release on secondary faults rather than the main thrust. In particular, the vertical motion expected at Middleton Island due to strain accumulation or slip on the main thrust is small because the dip angle is only about 3-4 degrees.

Brown *et al.* (1977) examined crustal uplift as indicated by tide gauge stations at Seward, Seldovia, St. Paul's Harbor (Kodiak), and Cordova as well as Anchorage, but their deduced rates are suspect due to the limited time span of the very noisy observations. A more robust set of tide gauge data was used by Savage and Plafker (1991) who analyzed sea level observations at seven southcentral Alaska sites (Anchorage, Kodiak Cordova, Seldovia, Seward, Nikiski, and Valdez) that were affected by the 1964 earthquake as well as five southeast Alaska sites (Juneau,

Ketchikan, Sitka, Skagway, and Yakutat) that were unaffected, and one southwest site (Sand Point) that was also unaffected. The rates of uplift they deduced for the southcentral Alaska locations are summarized on **Table 1**. Locales that subsided during the earthquake, notably Seldovia, Seward, and Nikiski showed postseismic uplift, whereas Cordova, which rose during the earthquake, showed postseismic subsidence. However, the quantitative correlation between the coseismic vertical displacement and the postseismic uplift rates was poor. Some of the postseismic uplift rates were particularly rapid. These include uplift rates of 17.7 ± 1.2 mm/yr at Kodiak, 9.9 ± 2.3 mm/yr at Seldovia, and 21.6 ± 2.0 mm/yr at Nikiski. The rate at Nikiski is suspect because the tide gauge was not operational from 1984 through 1996; however, the rates at Kodiak and Seldovia are based on more robust data sets. We will have more to say about these two rates, and their possible time dependence, shortly. Some of the other sites considered in the Savage and Plafker (1991) analysis showed relatively little motion. An example to which we referred earlier is Anchorage. Another is Seward where the post-1973 uplift rate was 0.0 ± 2.0 mm/yr. An important aspect of the analysis of Savage and Plafker (1991) is that the apparent annual sea level heights were adjusted by removing fluctuations that were correlated with those at sites in southeastern Alaska, a region not affected by the 1964 earthquake. The removal of an oceanographic signal resulted in a considerable reduction in the interannual noise in sea level height at some of the tide gauge locations (**Fig. 14**). Nevertheless, at Anchorage and to a lesser extent at Seward, there remained the previously mentioned oscillatory signal over the first few years following the earthquake. Cohen and Freymueller (2001) suggested the possibility that this oscillation might be an artifact of the data since some of the annual sea level heights used by Savage and Plafker (1991) were based on an incomplete record of monthly sea level heights and it is well known that the sea-level height has a strong annual signal. However, Larsen *et al.*

(2002) have shown that the oscillating signal remains when the analysis is based on monthly rather than annual sea level heights. Although the oscillation appears to be real, its origin is not known. Savage and Plafker (1991) suggested that it is tectonic rather than oceanographic, and Larsen *et al.* (2002) concurred, but neither provided a definitive mechanism for the behavior. If such an oscillation existed in the years following the earthquake then the leveling results along Turnagain Arm, which lies between Anchorage and Seward, might be contaminated by this unaccounted-for motion.

Savage and Plafker (1991) used tide gauge data available through 1988. Later, Cohen and Freymueller (2001) re-examined the tide gauge data making use of observations extending through another decade, i.e., until the end of 1998. The longer set of observations turned out to be critical for recognizing time-dependent changes in the vertical motion at the tide gauge sites. Even after the annual sea level heights were corrected for apparent oceanographic effects, the interannual noise was still quite large and a long set of observations was required in order to extract anything other than a linear trend from the data. Cohen and Freymueller (2001) found that there has been a very significant change in the uplift rate near Kodiak city. The mean postseismic uplift rate through the end of 1998 was 16.5 ± 0.7 mm/yr with the uplift rate changing at a rate of -0.67 ± 0.11 mm/yr² (**Fig. 15**) over the time span of the data. This rate change of nearly 7 mm/yr per decade brings other observed uplift rates in Kodiak in line with the tide gauge observations. These other observations include VLBI measurements from the late 1980's and early 1990's, recent GPS observations, and a combined analysis of daily tide gauge and Topex-Poseidon sea height observations (S. Nerem, private communication, 2001). From their tide gauge data analysis Cohen and Freymueller (2001) derived a time constant of about 12 years for a fifty percent reduction in the uplift rate from its peak. An unfortunate aspect of the Kodiak

observation history is that the tide gauge recording site was moved a short distance from St. Paul Harbor to Women's Bay in 1984. The repositioning of the tide gauge opens up the possibility that the apparent deceleration in the uplift rate was an artifact of the location change; but, this appears not to be the case. First, the two sites were connected by a leveling survey, then tide gauges were run at the two sites simultaneously for a couple months to verify that the height record was not distorted by the change. Thus, the fact that the time-averaged uplift rate at the St. Paul Harbor site (1967 to 1981) was 23.1 ± 1.4 mm/yr, while that from Women's Bay (1985-1998) was half that, 10.1 ± 1.1 mm/yr, should reflect the change in the uplift rate than the relocation of the tide gauge.

Cohen and Freymueller (2001) also found the subsidence rate has decreased at Cordova since the earthquake. The mean rate through 1998 was -4.7 ± 0.4 mm/yr while the rate change was 0.76 ± 0.14 mm/yr². About 17 years is required for a fifty percent reduction in the subsidence rate from its peak value. The confidence in the rate change at Cordova is not as strong as that for Kodiak because the rate change at Cordova is more dependent on the end points of the time series and there are fewer corroborating data. The rate change estimate did pass statistical testing (F-ratio test) for reliability, however. A further rate change, specifically at Valdez, is also hinted at by the data but the time series at Valdez is shorter than at Kodiak and Cordova, so the results are less definitive. The aforementioned analysis by Larson *et al.* (2002) also reveals a time dependence in the uplift rates at Kodiak, Cordova, and Valdez, but not at Nikiski and Seldovia.

The uplift rate derived by Cohen and Freymueller (2001) for Seldovia was a rather rapid 11.3 ± 0.8 mm/yr, a result consistent with that of Savage and Plafker (1991). While there was no indication of a time-dependence to this uplift rate, it is unlikely that a rate of about a centimeter per year can be sustained throughout the entire several-hundred-year recurrence period for a great

earthquake.

The uplift rate at Nikiski as deduced by Cohen and Freymueller (2001), 11.9 ± 0.8 mm/yr, was about half that derived by Savage and Plafker (1991), but the Nikiski data used by Savage and Plafker (1991) extended only into the early 1980's whereas Cohen and Freymueller (2001) also had data from 1997 and 1998.

It is intriguing to note that the tide gauges at the edge of the coseismic rupture zone, specifically those at Cordova and Kodiak appear to show rate decay times of a few decades. Shorter decay times were seen in the leveling data collected down-dip of the region of larger slip, i.e. along Turnagain Arm. No rate decay has been observed at Seldovia (probably because the decay time is longer than a few decades), a region of relatively small coseismic moment release. The speculation that the decay time is inversely related to the coseismic stress change is qualitatively consistent with the rheological properties of rocks.

Before leaving the discussion of tide gauge data, we should mention that most of the analyses have accounted for the effects of global sea level rise and postglacial rebound in only rudimentary fashion. Fortunately these two signals are estimated to produce not more than a few mm/yr of relative vertical motion, so detailed models have not been essential. Larsen *et al.* (2002) derived a simple model for vertical motions in Alaska based on data for past and present ice unloading (using, for the most part, data from Arendt *et al.*, 2002). This model suggests that the uplift from ice melting near the Kenai Peninsula is small. However, for some Earth models, uplift rates of up to several mm/yr occur on the Kenai Peninsula. These result mainly from the large ice unloading in southeast Alaska. This unloading could produce a broad scale uplift feature across the Kenai Peninsula; however, Larsen *et al.* (2002) find that the tide gauge data are too sparse to constrain the Earth model.

Some of the geodetic data sets collected in the aftermath of the earthquake have not been used to study crustal motion and others have been used only recently. The leveling profile from Matanuska to Glennallen apparently has not been systemically resurveyed since immediately after the earthquake. Indeed many of the survey markers from this profile may not have survived to the present. On the Kenai Peninsula the situation is better. A detailed leveling survey was conducted along the major roads of the Kenai Peninsula shortly after the earthquake. While this leveling survey was tied to that done along Turnagain Arm, no subsequent systematic releveing was performed on the Peninsula. In recent years, however, extensive Global Positioning System (GPS) surveys have been conducted in this region. Some of the GPS observations have been made at the surviving leveling marks, while others have been made at sites better suited for GPS techniques. These modern surveys began in 1993 and have been tied into other GPS surveys made along Turnagain Arm and those made near, and north of, Anchorage. Additional GPS observations have been made across Cook Inlet from the Kenai Peninsula as well as on Kodiak Island and near Valdez, although few, if any, of these observations are tied to historic leveling measurements. It should be noted, however, that even on the Kenai Peninsula, the majority of the historic survey markers are no longer present or cannot be found, having disappeared due to road construction or realignment and other causes.

Normally the comparison of heights obtained from leveling and GPS is difficult because the two heights are determined in fundamentally different coordinate system - the orthometric or sea level height system for leveling and the geometric coordinate system for GPS. However, these two height systems can be connected well enough to determine relative height changes if a high quality, local geoid height model is available (global geoid height models do not have the spatial resolution necessary to make the transformation). In the Kenai region, gravity observations were

made concurrent with the 1964 height observations. In addition, ship surveys in the nearby waters provide additional gravity information. These gravity data have been used by the US National Geodetic Survey (Smith and Milbert, 1999) to derive a detailed local geoid height model for Alaska as part of an overall effort to develop a precision geoid height model for the entire United States. The geoid height model has been used to determine relative height changes from a comparison of the recent GPS observations with the 1964 leveling data. In performing such an analysis one assumes that the time-dependent changes in the relative geoid height are small compared to the crustal movements; however, the largest source of error in the determination of height changes from this analysis is the uncertainty in this relative geoid height. This uncertainty increases with distance from the local reference. Cohen *et al.* (1995) and Cohen and Freymueller (1997) found that the maximum height change from 1964 to the early to mid 1990's relative to Seward was approximately 1 m. Seward is an appropriate reference locale since the tide gauge data indicate that it has very slow, if any, vertical movement. As shown in **Fig. 16**, the cumulative uplift since 1964 reaches a maximum in the center of the Kenai Peninsula and drops somewhat as one progress inland to the northwest. **Fig 16** is a revision of that presented in Cohen and Freymueller (1997) based on additional data and an updated geoid height model, GEOID99 (Smith and Roman, 2001). The updated model results in a greater consistency in the uplift for the sites in the western Kenai Peninsula. In addition to showing the uplift at leveling/GPS sites, the figure also shows the uplifts determined at several of the tide gauges discussed earlier. There is excellent agreement between the tide gauge and differenced GPS-leveling observations at Nikiski in the northwest portion of the Peninsula. The uplift derived from the tide gauge record at Seldovia appears to be somewhat less than that derived at the GPS-leveling sites observed just to the northeast, but the difference is not great and is consistent with

the expected fall-off in uplift with decreasing distance to the trench. Along Turnagain Arm the uplift exceeds 1 m at several locations. The uplifts along Turnagain Arm are derived, in part, by connecting leveling lines observed in 1964 on the Kenai Peninsula and along Turnagain Arm with the connection done at a site observed in both surveys. Thus, the height derived for the common site from least squares adjustment of the Kenai data was held fixed in the adjustment of the Turnagain Arm profile. Although the greater-than-1m uplift is consistent with the uplift derived on the Kenai Peninsula, the uplift would have to decrease rapidly to the northwest to be consistent with the nearly zero uplift at the Anchorage tide gauge location. Therefore, we cannot exclude the possibility that there is a significant bias in the uplift profile along the Turnagain Arm. More work is needed to determine whether the bias exists and, if it does, whether it is associated with an error in the geoid height model. Also the cumulative uplifts derived by comparing relatively recent observations with ones made shortly after the earthquake could be affected by oscillatory motions such as those observed at the Seward and Anchorage tide gauges if those oscillations were due to crustal movement. However, the peak cumulative uplift amplitudes well exceed any oscillation seen in the tide gauge signal.

VLBI observations conducted in the mid to late 1980's were the first systematic measurements of horizontal motion in southcentral Alaska made using space-geodetic techniques. Ma *et al.* (1990) deduced that the motion of the VLBI site on Kodiak Island, near Kodiak city, was 11.7 ± 1.0 mm/yr at an azimuth of $315^\circ \pm 5^\circ$ in a North America-fixed reference frame. They also deduced that Sourdough, a site located at $63^\circ 40'$ N, $146^\circ 29'$ W or somewhat less than 300 km north-northwest of Valdez, was moving not more than ~ 5 mm/yr relative to the interior of North America.

The advent of GPS technology has resulted in an extensive set of geodetic observations made

through the 1990's and into the 2000's. Freymueller *et al.* (2000) conducted GPS observations not only at the aforementioned leveling locales on the Kenai Peninsula but also a variety of newer sites better suited for GPS observations. Their observations have been combined with those of Savage *et al.* (1998) who examined a profile extending further toward the trench and further inland, but with less detail along-strike. Zweck *et al.* (2002a) presented an updated version of the velocity field. The pattern of horizontal crustal deformation is shown in **Fig. 17**. The figure shows an intriguing along-strike variation in the crustal motions, i.e. a difference between the velocities directions observed on the eastern and western sides of the Kenai Peninsula. The velocities on the eastern side displayed a pattern consistent with strain accumulation along a shallow locked plate interface. The velocities were oriented north-northwest consistent with the N17°W direction of the Pacific Plate relative to North America in this region [Zweck *et al.*, 2002a]. The velocities decreased with distance from the trench and were ~32 mm/yr near Seward. The exception to the general trend was a relatively low horizontal velocity of about 30 mm/yr at MIDD. This site is located about 90 km from the trench, presumably over a region where the subducting Pacific Plate is very shallow and not strongly coupled to the overriding North America Plate (MIDD also displayed an atypically large vertical velocity of ~20 mm/yr possibly due to continuous slip on a local fault branching upward from the megathrust). The maximum velocity observed was ~57 mm/yr at site MOTG, located ~150km from the trench. The velocity decreased to ~10 mm/yr at ~300 km from the trench.

Surprisingly the velocities of sites located on the western side of the Kenai Peninsula were oriented in almost opposite direction to those of the eastern side, indicating trenchward movement relative to stable North America. The magnitudes were significant, exceeding 20 mm/yr at several locations. The trenchward velocities of the western Kenai sites, unlike those on

the eastern side, are inconsistent with a model of a locked shallow plate boundary. This finding suggests that the coupling between the overriding North America Plate and subduction Pacific Plate is small in the western Kenai Peninsula, as opposed to the strong coupling nearer to Prince William Sound. The contrasting behavior between the eastern and western Kenai Peninsula velocities mirrors in some ways the differences in the coseismic moment release up-dip of the eastern and western Kenai Peninsula regions. As previously mentioned, there were large coseismic slips on the eastern Kenai Peninsula as part of the rupture of the Prince William Sound asperity. There is some evidence from seismic data that the length scale for this asperity was 140-200 km (Ruff and Kanamori, 1983). Between the eastern Kenai region and Kodiak Island, however, there was little moment release. Near Kodiak Island a second asperity ruptured during the 1964 earthquake, giving rise to significant coseismic slip and movement release. These observed velocity variations are purely empirical, whereas the inferences about coupling are somewhat model dependent. We will have more to say about the interpretation of the contemporary crustal movements observed in the vicinity of the Kenai Peninsula when we discuss the details of deformation modeling later in this article.

The contemporary crustal deformation across the Aleutian subduction zone near Kodiak has been studied by Savage *et al.* (1999) using GPS techniques. Over a distance from ~106 to ~250 km from the trench, observed velocities decrease with distance from the trench (see also Zweck *et al.*, 2002a). The velocities observed in this region are largely oriented in a direction consistent with plate convergence. The magnitudes, relative to stable North America, decrease from about 35 mm/yr for a site on Chirikof Island, southwest of Kodiak Island (see also, Fletcher *et al.*, 2001) to about 7 mm/yr for site KARL on the western side of Kodiak Island (**Fig. 18**). The elastic dislocation model of a locked subduction zone applied by Savage *et al.* (1999) to these data is

based on crustal movement relative to a site at Kodiak city. The more recent Zweck *et al.* (2002) model considers the velocities relative to the North America plate and includes some features not present in the earlier work. In particular, sites located 250 to 375 km from the trench on the Alaska Peninsula move trenchward, a behavior that is once again inconsistent with the model of a locked plate interface. In addition there is an along-strike variation in velocities of sites on the Pacific coast of Kodiak Island. Neither of these models explicitly consider the possibility that the deformation rates on Kodiak Island may be time dependent, although such a possibility is indicated by the aforementioned tide gauge observations.

Recent observations of crustal movement north of Anchorage have revealed another large-amplitude, transient phenomenon (**Fig. 19**), this one not being an immediate postseismic signal, but rather occurring several decades after the earthquake. The velocities at these sites as determined from GPS observations by Freymueller *et al.* (2001) for the period from about 1995 to late 1997 or early 1998 were consistent with other velocities on the eastern Kenai Peninsula. They were oriented in a north to north-northwest direction with magnitudes consistent with the decrease in crustal velocity with distance from the trench. However, beginning in 1998 and extending until late 2001, the velocities reversed direction and began to resemble those on the western Kenai Peninsula. The change in surface velocities had magnitudes as large as 25 mm/yr and were apparently due to slip on the plate interface at a depth of ~35 km. Freymueller *et al.* (2001) have called this phenomenon, “The Great Alaska ‘Earthquake’ of 1998-2001” because the rate and duration of the fault slip appears to be greater than that geodetically observed in similar events elsewhere. This event not only lasted longer than other recently observed subduction zone transients, it also differs in that it represents an acceleration of slip on a portion of the plate interface that was already slipping prior to the event.

Some words of caution are needed about comparing velocity plots provided by different workers and in comparing observed and computed velocities. In the several velocity plots shown above there are some ambiguities in the site locations relative to the trench. These ambiguities are associated with interpreting the location and strike of the trench. For example, Savage *et al.* (1998) plot the aforementioned MOTG site at 145-150 km from the trench, whereas Freymueller *et al.* (2000) plot the same site at about 170 km from the trench. Later, when we compare these data to models, another ambiguity arises because some of the models use a dip angle averaged over a considerable range of distances from the trench whereas the actual dip angle varies with location (albeit slowly in the shallow portions of the subduction zone). Thus, there is an uncertainty in the location of model distances relative to actual distance from the trench.

In addition to uncertainty in the location of various points relative to the trench and to model origins, there are also some systematic differences in the data analysis, particularly in the vertical motion. Freymueller *et al.* (2000) got consistently greater uplift rates than did Savage *et al.* (1998). The differences approached 10 mm/yr in some cases. For the most part these differences are due to reference frame issues and seem to have been largely resolved by careful consideration of these coordinate system matters. Recent work on vertical GPS velocities (Chris Larsen, private communication) show that Freymueller *et al.* (2000) overestimated vertical velocities because they used a reference frame tied to North America through site FAIR in Fairbanks rather than directly relating International Terrestrial Reference Frame velocities to North America. As is typically the case, there is far better consistency in the horizontal velocities as reported by various researchers.

4. Models of Southcentral Alaska Preseismic and Postseismic Crustal Deformation

In the preceding sections the focus of discussion has been on the geodetic measurements themselves and direct inferences from those measurements that are not highly dependent on specific model assumptions. This is not to say that model assumptions have not entered either implicitly or explicitly into the results we reviewed. Indeed, a variety of geophysical models are often required in order to reduce the raw observations of geodetic data. Many of these models are needed to convert that actual measurement observables, such as the time a signal is received from a GPS satellite or the height observed on a measuring rod, into geodetic quantities such as three-dimensional positions. Examples include models of solid earth tides, atmospheric and ionospheric propagation, relativistic effects, and various forcings on the orbits of the GPS satellites. Other models are needed to infer physical significant parameters from the geodetic quantities. For example, determinations of coseismic slip often require a model of the subduction zone geometry. This geometry is often constrained from seismic observations, which in themselves may depend on other model assumptions.

In the present section our focus is not on the models used for data reduction, but rather on those that have been used to interpret the geodetic data. We have already commented on the importance that models of the coseismic motion had in the development of ideas that shallow thrust, subduction zone earthquakes are a key component of plate tectonics. Various models based on elastic rebound theory have been exploited to explain both not only coseismic, but also postseismic and interseismic motion. Attempts have been made (with not entirely satisfactory results) to combine elastic rebound theory with plate bending theory as in Rosenbaum (1974).. An important point to remember when considering both the models of crustal deformation and the data they seek to explain is that coseismic and interseismic deformations are often oppositely directed and of nearly equal magnitude when considered over several or even a single earthquake

cycle in a subduction zone. Since long-term geologic deformation is the cumulative result of these and other motions (such as glacial rebound) it usually develops at a much slower rate than short term indicators would suggest.

In modeling strain accumulation, the most commonly employed model is that of a locked fault plane embedded in either an elastic half-space or layered elastic-viscoelastic half-space. A conventional method for modeling the accumulation of strain is to conceptually decompose the driving motion into two terms associated with slip along the plate interface (Savage and Prescott, 1978; Savage, 1983). The first term is due to steady-state, i.e. constant velocity, fault-slip, and gives rise to nearly rigid block translation with very slow or no stress or strain accumulation. The second term is cyclic and consists of two parts: the response to imagined backward slip along the locked portion of the megathrust between earthquakes and the response to forward slip (usually along the same segment of the megathrust) during the earthquake. This cyclic term is the driver of crustal deformation. Although this conceptual decomposition has been applied to both elastic and viscoelastic models, it is particularly simple in the former case because there are analytical solutions for different types of dislocations embedded in an elastic half-space. The simple model is often made more sophisticated by considering complexities in the fault geometry, the finite thickness of the subducting and overriding plates, and strain accumulation and/or slip on subsidiary faults. Several examples of elastic dislocation calculations, as applied to geodetic data observed in Alaska, will be presented below.

Two competing models, one involving deep fault creep and the other viscoelastic flow, are often investigated as mechanisms for transient postseismic deformation. The differences between the physics of these two models are discussed in considerable detail in Cohen (1999). Here we just summarize a few salient features. When an earthquake ruptures through a finite portion

of the lithosphere, there is, of course, a sudden change in the stress field in the surrounding medium. In particular the release of stress in the upper layers of the crust and mantle is accompanied by a sudden increase in stress below that. There are several possible responses of the Earth to this stress redistribution. One common response is for a portion of the fault plane lying downdip of the coseismic rupture to slip aseismically in order to relieve the stress imposed across it. Typically, the surface deformation due to this process is computed by considering slip on a buried fault in an elastic half-space, although the half-space does not necessarily have spatially uniform elastic properties. Another response is for ductile flow to occur in portions of the lower crust and/or mantle, with the depth of this flow depending on temperature, pressure, and rock type. The ductile flow couples back to the overlying elastic layer, causing both surface and subsurface deformations. In postseismic rebound studies the ductile flow is usually modeled as a viscoelastic process with the strain rate proportional to stress (linear viscoelastic flow) or a power of stress (non-linear, power law flow). The viscous properties are temperature-dependent and therefore can vary greatly with depth, although the majority of the models considered so far use a single value for viscosity. Creep at depth and viscoelastic flow are not the only processes that can produce significant postseismic deformation. Some other candidates include surface aftercreep and poroelasticity. In addition, the complex fault rheology implied by state-dependent friction laws can produce temporal and spatially varying crustal deformation through both stable and unstable slip responses to stressing.

In trying to interpret the uplift observed along the Turnagain Arm, Brown *et al.* (1977) considered creep on a buried fault, viscoelastic flow, strain accumulation on the megathrust due to the continuing convergence between the North America and Pacific Plates, strain accumulation on a subsidiary fault, subsurface magmatic activity and dilatancy effects. They

concluded that the latter two mechanisms did not offer a viable explanation for the observed uplift, but that buried slip, viscoelastic rebound, and strain accumulation on the megathrust were realistic possibilities. Given the rate of plate convergence it is unlikely that strain accumulation, by itself, could have explained the bulk of the postseismic uplift and it certainly did not provide an explanation for the time dependence of the uplift. However as Brown *et al.* (1977) found, down-dip creep on the megathrust amounting to a couple meters slip between 1964 and 1975 did offer a satisfactory explanation for the observed uplift. At least part of deep slip region might be on that portion of the plate interface that did not rupture during the earthquake, but lies above the aforementioned thermally allowed downdip limit to the locking depth. Viscoelastic flow could not be entirely excluded as an alternative explanation. However, this mechanism was less satisfactory to Brown *et al.* (1977) because they could not find supporting evidence in seismic data for a viscosity small enough to explain the ~5 yr relaxation time observed in the uplift rates.

In contrast to Brown *et al.* (1977), Wahr and Wyss (1980) argued that a viscoelastic model did provide a good explanation for the uplift observed after three Alaska earthquakes: not only the 1964 earthquake, but also an event in the central Aleutians in 1957 and another in the western Aleutians in 1965. They picked a Maxwell relaxation time, $\tau = \eta/\mu = 1\text{yr}$, where η is the viscosity of the viscoelastic material and μ the material's rigidity. The novel aspect of their model is that the viscoelastic material was not placed in a horizontal layer, the asthenosphere, but rather in a rectangular inclusion in the overthrust block adjacent to an unlocked portion of the plate interface. In the case of the 1964 earthquake, the inclusion was taken be 40 km wide and 60 km high extending from depths of 20 to 80 km. The required inclusion appears to be more than 400 km from the trench and would lie about 25 km north of Anchorage. One objection to this model is that the location of the inclusion does not coincide with existing volcanic features. A

somewhat lesser difficulty is that a Maxwell time of 1 yr implies a very low viscosity of about 10^{18} Pa s. Such a low value seems unlikely but could, perhaps, be reconciled with rock mechanics if the effects of a power law rheology and the large stress change associated with an earthquake are considered or flow is governed by the transient response of a composite viscous material in the lower crust (Ivins, 1996). Cohen (1996) considered both elastic afterslip, viscoelastic rebound in a low viscosity layer, and a combination of the two processes as a mechanism for the observed uplift on the Kenai Peninsula. The simplest model that was consistent with the data was a pure afterslip model. By contrast, a purely viscoelastic model was not very successful in that the model predictions either: (1) failed to fit the contemporary observations or (2) gave unrealistically large uplifts later in the earthquake cycle. A combined model involving both afterslip and viscoelastic flow did provide an adequate explanation for the data. The required afterslip was greater than that required in the purely elastic model because the slip had to be sufficient both to compensate for the subsidence produce by viscoelastic flow and to produce the observed vertical upward motion (**Fig. 20**)

Lundren *et al.* (1995) published a finite element, elastic, spherical shell model of crustal deformation in Alaska. The boundary conditions and constraints they imposed on the model were selected fault orientations and slip rates and baseline rates of change derived from VLBI observations at Nome, Fairbanks (Gilcreek), Sourdough, Alaska and Whitehorse, Canada. Many of the parameters of their model would probably be revised if the analysis were to be repeated today since contemporary geodetic observations (e.g. Fletcher and Freymueller, 1999), as well as new geological information, have provided a wealth of constraints not available to the earlier workers. However, the study was noteworthy as one of earlier attempts to integrate space geodetic data and geologic data into a mutually consistent model of crustal deformation in

Alaska.

Piersanti *et al.* (1997) considered the effects of viscoelastic rebound to the 1964 earthquake on VLBI measurements between a site near Fairbanks in central Alaska and Whitehorse in northwestern Canada as well as with several other sites in Alaska, specifically Nome on the west coast, Kodiak Island in the southcentral part of the state, near the 1964 rupture zone, Sand Point, in the central Aleutians west of the rupture, Yakataga, in southcentral Alaska, east of the rupture zone, and at Sourdough, also in the interior of Alaska. Their model employed a spherical Earth geometry and a 200 km low viscosity channel with viscosities in the range 10^{19} - 10^{20} Pa s. The computed surface velocities, both along the baseline between the sites and transverse to it, for the mid-to-late 1980 time-frame ranged from 0.1 mm/yr to as much as 9 mm/yr. The observed velocities were generally of the same magnitude, but overall the agreement between the predicted and observed velocities was not particularly good. The model calculations assumed a dip angle for the coseismic rupture plane of 20° , much too steep a dip. The fault was also placed at much greater depths than indicated by the most reliable studies of the PWS earthquake. Other model parameters are also somewhat questionable in view of what is now known both about the geometry of the interface and the parameters of the earthquake.

The contemporary horizontal crustal velocities observed in the vicinity of the Kenai Peninsula and Kodiak Island were discussed in the previous section. The northwesterly oriented crustal velocities on the eastern portions of the Kenai Peninsula and in the Kodiak-Katmai region have been analyzed using conventional elastic dislocation models of strain accumulation on the shallow locked portion of the plate interface. For example, Freymueller *et al.* (2000) experimented with several different models. In the first, the locked plate interface was represented by a single plane. In this model the dip angle was set to 3° , a value derived from

seismological observations. The adjusted parameters of the model are the down-dip end of the locked interface and the back-slip velocity. In the second model, the dip-angle was also adjusted. In the third model, postseismic creep was also permitted on a deep segment of the interface. The parameters of the three models are given in Table 2 and the comparison between observed and computed velocities for the models is shown in **Fig. 21**. Taken as a whole the suite of models provide a satisfactory fit to the horizontal velocity data, but fail to account adequately for the vertical data.

Savage *et al.* (1998) also used an elastic dislocation model to explain the northwestward motion of the sites observed across the central portion of the 1964 rupture zone (**Fig. 22**). They used a back-slip velocity of 65 mm/yr instead of the plate convergence rate of about 55 mm/yr in an attempt to account for a long-term transient response to the earthquake; however, it is not clear to what extent the higher than plate velocity convergence rate would be required in a full three dimensional model that also takes into account the very shallow dip angle of the plate interface at shallow depths

As mentioned earlier, the trenchward motion of sites on the western Kenai Peninsula cannot be explained by any simple model of strain accumulation along a locked plate interface. Freymueller *et al.* (2000) suggested that the difference between the north-northwest motion of the eastern side of the Kenai Peninsula and the southwest motion on the west (relative to North America) is a reflection of short wavelength variability in plate coupling with strong coupling in the east and weak coupling in the west. An intriguing possibility is that the coupling is not only spatially, but temporally varying and reflects a diffusive or propagating phenomenon at depth.

Zweck *et al.* (2002a) extended the previous two-dimensional inversion analysis of the GPS data into three dimensions and included a larger GPS data set. They derived a model of the coupling

distribution between the Pacific and North America plate by inverting crustal velocities for over 200 GPS sites in southcentral Alaska, with most of them being located on or near the Kenai Peninsula, Prince William Sound, Kodiak Island and portions of the Alaska Peninsula. In agreement with comments just made about the spatial variability in coupling, they found the plates to be strongly coupled at the Prince William Sound and Kodiak Island asperities (**Fig. 23**) and suggested that delayed or residual postseismic reverse slip at a rate of 80-100 mm/yr at depths of 30-40 km is responsible for the observed trenchward motion on the western side of the Kenai Peninsula. The same mechanism may also be responsible for trenchward velocities on the Alaska Peninsula near Kodiak Island. Savage *et al.* (1999) inferred that the shallow plate interface beneath and offshore of southwestern Kodiak Island is strongly coupled. However, as we mentioned earlier, this analysis of relative velocities misses the long wavelength impact of the postseismic deformation. By contrast, Zweck *et al.* (2002a) considered the deformation field over a broad area including Kodiak Island and most of southcentral Alaska. They deduced that the trenchward motion of sites arcward of Kodiak city, and the along-strike variations, are due mainly to continuing postseismic deformation following the 1964 earthquake.

Zweck *et al.* (2002b) used an inversion analysis of present day GPS, 35-year averaged leveling, and triangulation data to estimate back-slip rates and coupling in the western portion of the Kenai Peninsula over two different time periods. By comparing present day deformations with cumulative deformations since the earthquake, they concluded that average back-slip since the earthquake was about 60 mm/yr or 3-4 times larger than the contemporary rate. They also found that the temporal decay in the rate of postseismic slip cannot be explained by a single time-decaying process with either an exponential or logarithmic time dependence. They found, however, that the temporal decay might be explained by a combination of the two constants, or

by a slip source that changes in location and slip velocity over time.

Bird (1996) developed a two-dimensional finite element model of the Alaska-Bering Sea region that included variable crust and lithospheric thickness, heat flow, and many plate boundary and interior faults. He investigated 46 variants of the model to investigate the effects of shear traction on the megathrust, fault friction, internal friction of lithospheric blocks and mantle creep strength. He concluded that both fault friction and driving traction on the Aleutian megathrust are low. He also concluded that the present neotectonic regime is a transient response to Plio-Pleistocene glacial mass redistribution and differs from typical tectonic flow in the tertiary. His model predictions cannot be tested against southcentral Alaska geodetic observations because he considers only long term deformation rather than the earthquake cycle.

5 Summary

Some of the specific observations of crustal deformation in southcentral Alaska are, in themselves, remarkable. Examples include: ~25m of horizontal coseismic motion, ~150 mm/yr relative uplift in the year after the earthquake, and over 1 m cumulative uplift in the 40 years since the earthquake. Progressing from east to west across the Kenai Peninsula there are ~180° rotations in the contemporary velocity vectors. There are temporal reversals in the velocity vector orientations near Anchorage in the late 1990s. The overarching theme that emerges from both observational and modeling studies of crustal deformation for the PWS earthquake zone is one of temporal and spatial variability. Although there is much still to be learned about the mechanisms responsible for the time-dependent crustal deformation signals, it seems likely that a variety of processes interact to produce the complex signal that is observed. These processes may include fault-centered stable and unstable sliding phenomena, as governed by state and velocity

dependent friction laws (e.g. Test and Rice, 1986), and bulk phenomena such as viscoelastic flow. As geodetic observations in southcentral Alaska span only a few decades it is quite remarkable that a myriad of deformation styles have already been observed. However, the Alaska seismic zone is not unique in revealing a complexity in the temporal-spatial variation in crustal deformation. For example, silent earthquakes and aseismic slip events have been recently observed at subduction zones in Cascadia, Japan, and South America. Although the deformations that have been observed so far in southcentral Alaska are due primarily to plate tectonic and seismic processes, it is likely that the more subtle effects of glacial rebound and long-term, long-wavelength relaxation processes will have to be considered in future studies.. Models of increasing sophistication that can account both for earthquake-cycle and permanent deformation will be required in order to explain the seismotectonics and geological evolution of the subduction zone.

Acknowledgements: The authors have worked with many colleagues whose have willing shared their insights and observations on Alaska seismotectonics. We express our appreciation, in particular, for the many contributions of Chris Larsen, Hilary Fletcher, Sandy Holdahl, Jeanne Sauber, Max Wyss, and Chris Zweck.

6. References

- Arendt, A. A., Echelmeyer, K. A., Harrison, W. D., Lingle, C. S., and Valentine, V. B. (2002). Rapid wastage of Alaska glaciers and their contribution to rising sea level, *Science*, **297**, 392-286.
- Atwater, T., and Severinghouse, J. (1989). Tectonic maps of the northeast Pacific. In "The eastern Pacific Ocean and Hawaii," Winterer, E. L., Hussong, D. M. and Decker, R. W., eds., The Geology of North America, Geological Society of America, Boulder, pp 15-20.
- Bartsch-Winkler, S., and Schmoll, H. R. (1992). Utility of radiocarbon-dated stratigraphy in determining late Holocene earthquake recurrence intervals, upper Cook Inlet region, Alaska, *Geol. Soc. Am. Bull.*, **104**, 684-694.
- Beckman, H. M. (1980). Geologic Map of Alaska, U.S. Geological Survey Special Map.
- Bird, P. (1966). Computer simulations of Alaskan neotectonics, *Tectonics*, **15**, 225-236.
- Brown, L. D., Reilinger, R. E., Holdahl, S. R., and Balazs, E. I. (1977). Postseismic crustal uplift near Anchorage, Alaska. *J. Geophys. Res.* **92**, 3369-3377.
- Brocher, T. M., Fuis, G. S., Fisher, M. A., Plafker, G., Moses, M. J., Taber, J. J., and Christensen, N. I. (1994). Mapping the megathrust beneath the northern Gulf of Alaska using wide-angle seismic data. *J. Geophys. Res.* **99**, 11663-11685.
- Carver, G. A., and McCalpin, J. P (1996) Paleoseismology of compressional tectonic environments. In Paleoseismology, pp 231-232 and pp 183-270, McCalpin, J., ed., Academic Press, San Diego.
- Christensen, D. H., and Beck, S .L. (1994). The rupture process and tectonic implications of the great 1964 Prince William Sound earthquake. *Pure Appl. Geophys.* **142**, 9-53.

- Cohen, S. C. (1996). Time-dependent uplift of the Kenai Peninsula and adjacent areas of south central Alaska since the 1964 Prince William Sound earthquake. *J. Geophys. Res.* **101**, 8595-8604.
- Cohen, S. C. (1998). On the rapid postseismic uplift along Turnagain Arm, Alaska following the 1964 Prince William Sound earthquake. *Geophys. Res. Lett.* **28**, 1213-1215.
- Cohen, S. C. (1999). Numerical models of crustal deformation in seismic zones. In "Advances in Geophysics **41**, (Dmowska, R. and Saltzman, B., eds.)," Academic Press, San Diego, pp 133-232.
- Cohen, S. C., Holdahl, S., Caprette, D., Hilla, S., Safford, R., and Schultz, D. (1995). Uplift of the Kenai Peninsula, Alaska, since the 1964 Prince William Sound earthquake. *J. Geophys. Res.* **100**, 2031-2038.
- Cohen, S. C., and Freymueller, J. T. (1997). Deformation on the Kenai Peninsula, Alaska. *J. Geophys. Res.* **102**, 20479-20487.
- Cohen, S. C., and Freymueller, J. T. (2001). Crustal uplift in the south central Alaska subduction zone: New analysis and interpretation of tide gauge observations. *J. Geophys. Res.* **106**, 11259-11270.
- Combellick, R. A. (1991). Paleoseismicity of the Cook Inlet Region, Alaska: Evidence from peat stratigraphy in Turnagain and Knik Arms, Professional Report 112, State of Alaska, Division of Geological and Geophysical Surveys, Fairbanks.
- Combellick, R. A. (1992). The penultimate great earthquake in southcentral Alaska: Evidence from a buried forest near Girdwood. Public Data File 92-12, State of Alaska, Division of Geological and Geophysical Surveys, Fairbanks.
- Combellick, R. A. (1993). Investigation of peat stratigraphy in tidal marshes along Cook Inlet,

- Alaska to determine the frequency of 1964-style great earthquakes in the Anchorage region. Public Data File 93-84, State of Alaska, Division of Geological and Geophysical Survey, Fairbanks.
- DeMets, C., Gordon, R. G., Argus, D. F. and Stein, S. (1990). Current plate motions. *Geophys. J. Int.* **101**, 425-478.
- DeMets, C., Gordon, R. G., Argus, D. F. and Stein, S. (1994). Effects of recent revisions to the geomagnetic reversal time scale on estimates of current plate motions. *Geophys. Res. Lett.*, **21**, 2191-2194.
- Demets, C. and Dixon, T. H. (1999). New kinematic models for Pacific-North America motion from 3 MA to present, 1: Evidence for steady motion and biases in the NUVEL-1A model. *Geophys. Res. Lett.*, **26**, 1921-1924.
- Doser, D. I., Veileux, A. M., and Velasquex, M. (1999). Seismicity of the Prince William Sound region for over thirty years following the 1964 great Alaska earthquake. *Pure Appl. Geophys.* **154**, 593-632.
- Estabrook C. H., and Jacob, K. H. (1991) Stress indicators in Alaska. In "Neotectonics of North America, Decade Map Volume 1, Chapter 21," Slemmons, D. B., Engdahl, D. B., Zoback, M. D., and Blackwell, D. D. eds., Geological Society of America, Boulder.
- Estabrook, C. H., Jacob, K. H., and Sykes, L. R. (1994). Body wave and surface wave analysis of large and great earthquake along the Eastern Aleutian Arc, 1923-1993: Implications for future events. *J. Geophys. Res.*, **99**, 11643-11662.
- Fletcher, H. J., and Freymueller, J. (1999). GPS constraints on the motion of the Yakutat Block. *Geophys. Res. Lett.*, **26**, 3029-3032.
- Fletcher, H.J., and Freymueller, J. (2002). Tectonics of interior Alaska, A study using GPS

- measurements. American Association for the Advancement of Science (abstract).
- Fletcher, H. J., Beavan, J., Freymueller, J., and Gilbert, L. (2001). High interseismic coupling of the Alaska subduction zone SW of Kodiak island inferred from GPS data. *Geophys. Res. Lett.*, **28**, 443-446.
- Freymueller, J. T., Cohen, S. C., and Fletcher, H. (2000). Variations in present day deformation, Kenai Peninsula, Alaska, and their implications. *J. Geophys. Res.* **105**, 8079-8102.
- Freymueller, J. T., Zweck, C., Fletcher, H., Hreinsdottir, S., Cohen, S. C., and Wyss, W. (2001). The great Alaska "earthquake" of 1998-2001. *Eos, Trans. Am. Geophys. U.* 82(47), Fall Meet. Suppl. Abstract..
- Harding, S. T., and S T. Algermissen (1972). The focal mechanism of this and related earthquakes. In "The Great Alaska Earthquake of 1964 , Seismology and Geodesy," National Academy of Sciences, Washington, D.C., 197-233.
- Hastie, L. M., and J. C. Savage (1970). A dislocation model for the 1964 Alaska earthquake. *Bull. Seismol. Soc. Am.*, **60**, 1389-1392, 1970.
- Holdahl, S. R., and Sauber, J. (1994). Coseismic slip in the 1964 Prince Williams Sound earthquake: A new geodetic inversion. *Pure Appl. Geophys.* **142**, 55-82.
- Ivins, E. R. (1996). Transient creep of a composite lower crust 2. A polymineralic basis for rapidly evolving postseismic deformation modes, *J. Geophys. Res.*, **101**, 28005-28028.
- Johnson, J. M. (1999). Heterogeneous coupling along the Alaska-Aleutians as inferred from Tsunami, seismic, and geodetic inversions. In "Advances in Geophysics," **39**, (Dmowska, R. and Saltzman, B., eds.), Academic Press, San Diego, pp 1-116.
- Johnson, J. M., Satake, K., Holdahl, S. R., and Sauber, J. (1996). The 1964 Prince William Sound Earthquake: Joint inversion of tsunami and geodetic data. *J. Geophys. Res.*, **101**, 522-

532.

- Kanamori, H., (1970). The Alaska earthquake of 1964: Radiation of long-period surface waves and source mechanisms. *J. Geophys. Res.*, **75**, 5029-5040.
- Kikuchi, M., and Fukao, Y. (1987). Inversion of long-period P waves from great earthquakes along subduction zones. *Tectonophysics*, **144**, 231-247.
- Larsen, C., Motyka, R. Freymueller, J. T., and Echelmeyer, K. A.. (2002). Tide gauge records of uplift along the northern Pacific – North America plate boundary, 1937 to 2001. *J. Geophys. Res.*, (submitted).
- Lay, T., and Wallace, T. C. (1995). *Modern Global Seismology*, pp. 7, Academic Press, San Diego.
- Little, T. A. and C. W. Naeser, (1989). Tertiary tectonics of the Border Ranges Fault System, Chugach Mountains, Alaska: Deformation and uplift in a forearc setting. *J. Geophys. Res.*, **94**, 4333-4359.
- Lundgren, P., Saucier, F., Palmer, R., and Langon, M. (1995). Alaska crustal deformation: Finite element modeling constrained by geological and very long baseline interferometry data. *J. Geophys. Res.* **100**, 22033-22045.
- Ma, C., Sauber, J. M., Bell, L. J., Clark, T. A., Gordon, D., Himwich, E., and Ryan, J. W. (1990). Measurement of horizontal motions in Alaska using very long baseline interferometry. *J. Geophys. Res.* **95**, 21991-22011.
- Miyashita, K., and Matsu'ura, M. (1978). Inversion analysis of static displacement data associated with the Alaska earthquake of 1964. *J. Phys. Earth* **26**, 333-349.
- Moore, J. C. , Diebold, J., Fisher, M. A., Sample, J., Brocher, T., von Huene, R., Rowe, C., Stone, D., Stevens, C., and Sawyer, D. (1991). EDGE deep seismic reflections transect of the

eastern Aleutian arc-trench layered lower crust reveals underplating and continental growth.

Geology, **19**, 420-425.

Nishenko, S. P., and Jacob, K. H. (1990). Seismic potential of the Queen Charlotte-Alaska-Aleutian seismic zone. *J. Geophys. Res.* **95**, 2511-2532.

Oldow, J. S., Bally, A. W., Ave' Lallemant, H. G., and W. P. Leeman (1989). Phanerozoic evolution of the North American cordillera; United States and Canada. In *The Geology of North America- An overview* (Bally, A. W., and R. A. Palmer, eds.), Geologic Society of America, pp139-232.

Oleskevich, D. A., Hyndman, R. D., and Wang, K. (1999). The updip and downdip limits to great subduction earthquakes: Thermal and structural models of Cascadia, south Alaska, SW Japan, and Chile. *J. Geophys. Res.* **104**, 14965-14991.

Page, R. A., Biswas, N. N., Lahr, J. C., and Puplan, H. (1991). Seismicity of continental Alaska. In "The Geology of North America, Decade Map Vol. 1", The Geological Society of America, pp. 47-68.

Parkin, E. J. (1972). Horizontal crustal movements. In "The Great Alaska Earthquake of 1964 , Seismology and Geodesy," National Academy of Sciences, Washington, D.C., pp 419-434.

Pavlis, T. L., and G. W. Crouse (1989). Late Mesozoic strike slip movement on the Border Ranges Fault System in the Eastern Chugach Mountains, Southern Alaska. *J. Geophys. Res.*, **94**, 4321-4332.

Perez, O. J., and Jacob, K., H. (1980). Tectonic model and seismic potential of the eastern Gulf of Alaska and Yakataga seismic gap. *J. Geophys. Res.* **85**, 7132-7150.

Piersanti, A., Spada, G., and Sabadini, R. (1997). Global postseismic rebound of a viscoelastic Earth: Theory for finite faults and application to the 1964 Alaska earthquakes. *J. Geophys. Res.*

102, 477-492.

Plafker, G. (1965) Tectonic deformation associated with the 1964 Alaskan earthquake, *Science*, **148**, 1675-1687.

Plafker, G. (1972). Tectonics. In "The Great Alaska Earthquake of 1964 , Seismology and Geodesy," National Academy of Sciences, Washington, D.C., 113-188.

Plafker, G., Lajoie, K. R., and Rubin, M. (1992). Determining recurrence intervals of great subduction zone earthquakes in southern Alaska by radiocarbon dating. In "Radiocarbon After Four Decades: An Interdisciplinary Perspective" (Taylor, R. E., Long, A., and Kra, R. S., eds.), New York, Springer-Verlag, p. 436-453.

Plafker, G. Gilpin, L., and Lahr, J. (1994a) Neotectonic map of Alaska (1993). In "The Geology of North America, Vol. G-1, the Geology of Alaska," The Geological Society of America, pp. 389-449.

Plafker, G. Moore, J. C. and Winkler, G. R. (1994b). Geology of the southern Alaska margin, In "The Geology of North America, Vol. G-1, the Geology of Alaska," The Geological Society of America, pp. 389-449.

Ponko, S. C., and Peacock, S. M. (1995). Thermal modeling of the southern Alaska subduction zone: Insight into the petrology of the subducting slab and overlying mantle wedge. *J. Geophys. Res.*, **128**, 22117-22128.

Pope, A. J. (1972). Strain analysis of horizontal crustal movements in Alaska based on triangulation surveys before and after the Earthquake. In "The Great Alaska Earthquake of 1964 , Seismology and Geodesy," National Academy of Sciences, Washington, D.C., pp435-448.

Prescott, W. H., and Lisowski, M. (1977). Deformation at Middleton Island, Alaska during the decade after the Alaska earthquake of 1964. *Bull. Seismo. Soc. Am.* **67**, 579-586.

- Prescott, W. H., and Lisowski, M. (1980). Deformation at Middleton Island, Alaska. *Bull. Seismo. Soc. Am.* **70**, 1887-1892.
- Press, F. (1965). Displacements, strains, and tilts at teleseismic distance. *J. Geophys. Res.* **70**, 2395-2412.
- Press, F., and Jackson, D. J., (1965). Vertical extent of faulting and elastic strain energy release. *Science*, **147**, 867-868.
- Pulpan, H., and Frohlich, C. (1985). Geometry of the subducted plate near Kodiak Island and lower Cook Inlet, Alaska, determined from relocated earthquake hypocenters. *Bull. Seismo. Soc. Amer.*, **75**, 791-810.
- Rosenbaum, J. G., (1974). Theory of warping of southern Alaska before Alaska earthquake, 1964. *J. Geophys. Res.* **79**, 3294-3301.
- Ruff, L., and Kanamori, H. (1983). The rupture process and asperity distribution of three great earthquakes from long-period diffracted P-waves. *Phys. Earth Planet. Int.*, **31**, 202-229.
- Rundle, J. B., Turcotte, D. L., and Klein, W. (2000). Introduction in "Geocomplexity and the Physics of Earthquakes", American Geophysical Union Geophysical Monograph 120, Washington, DC.
- Savage, J. C. (1983). A dislocation model of strain accumulation and release at a subduction zone. *J. Geophys. Res.*, **88**, 4984-4996.
- Savage, J. C., and Prescott, W. H. (1978). Asthenospheric readjustment and the earthquake cycle, *J. Geophys. Res.*, **83**, 3369-3376.
- Savage, J. C., and Lisowski, M. (1991). Strain accumulation along the Denali Fault at the Nenana River and Delta River Crossings, Alaska. *J. Geophys. Res.*, **96**, 14481-14492.
- Savage, J. C., and Plafker, G. (1991). Tide gage measurements of uplift along the south coast of

- Alaska. *J. Geophys. Res.* **96**, 4325-4335.
- Savage, J. C., Svarc, J. L., Prescott, W. H., and Gross, W. K. (1998). Deformation across the rupture zone of the 1964 Alaska earthquake, 1993-1997. *J. Geophys. Res.*, **103**, 21275-21283.
- Savage, J. C., Svarc, J. L., and Prescott, W. H. (1999). Deformation across the Alaska-Aleutian subduction zone near Kodiak. *Geophys. Res. Lett.* **26**, 2117-2120.
- Sella, G. F. Dixon, T. H. and Mao, A. (2002) REVEL: A model for recent plate velocities from space geodesy. *J. Geophys. Res.*, **107**, 10.1029/2000JB000033.
- Shennan, I. A., Scott, D. B., Rutherford, M., and Zong, V. (1999). Microfossil analysis of sediments representing the 1964 earthquake, exposed at Girdwood Flats, Alaska, USA. *Quaternary International*, **60**, 55-73.
- Sherburne, R. W., Algermissen, S.T., and Harding, S.T. (1972). Hypocenter, origin time, and Magnitude, in "The Great Alaska Earthquake of 1964 , Seismology and Geodesy," National Academy of Sciences, Washington, D.C., pp. 29-49.
- Small, J. B. and Wharton, L. C. (1972). Vertical displacements. In "The Great Alaska Earthquake of 1964 , Seismology and Geodesy," National Academy of Sciences, Washington, D.C., pp. 449-461.
- Smith, D. A., and Milbert, D. G. (1999) The GEOID96 high resolution geoid height model for the United States. *J. Geod.*, **73**, 219-236.
- Smith, D. A., and Roman, D. R. (2001) GEOID99 and G99SSS: 1-arc-minute geoid models for the United States. *J. Geod.*, **75**, 469-490.
- Stauder, W. (1972). Introduction (to I: Parameters of the main shock). In "The Great Alaska Earthquake of 1964 , Seismology and Geodesy," National Academy of Sciences, Washington, D.C., pp. 23-28.

- Stephens, C. D., Page, R. A., and Lahr, J. C. (1990). Reflected and mode-converted seismic waves within the shallow Aleutian subduction zone, southern Kenai Peninsula, Alaska. *J. Geophys. Res.* **95**, 6883-6897.
- Stauder, W., and Bollinger, G. A. (1966). The focal mechanism of the Alaska earthquake of March 28, 1964 and of its aftershock sequence. *J. Geophys. Res.*, **71**, 5283-5295).
- Tichelaar, B. W. and Ruff, L. J. (1993). Depth of seismic coupling along subduction zones. *J. Geophys. Res.*, **98**, 2017-2037.
- Tocher, D. (1972). General Introduction: Tectonics and seismic effects of the Alaska earthquake and the seismicity of Alaska, In "The Great Alaska Earthquake of 1964 , Seismology and Geodesy," National Academy of Sciences, Washington, D.C., pp. 1-7.
- Test, S. T. and Rice, J. R. (1986). Crustal earthquake instability in relation to depth variation of frictional properties. *J. Geophys. Res.*, **91**, 9452-9472.
- Udike, R. G., and Schmoll, H. R. (1985). A brief resume' of the geology of Anchorage and vicinity, Public-data File 85-2, Alaska Division of Geological and Geophysical Surveys, Fairbanks.
- Vine, F. J., and Mathews, D. H. (1963). Magnetic anomalies over oceanic ridges. *Nature*, **199**, 947-949.
- Von Huene, R., Fisher, M. A., and Bruns, T. R. (1987). Geology and evolution of the Kodiak margin, Gulf of Alaska. In "*Geology and resource potential of the continental margin of western North America and Adjacent Ocean Basins – Beaufort Sea to Baja California*," (Scholl, D. W., Grantz, A., and Vedder, eds.) 1987. Texas Circum-Pacific Council for Energy and Mineral Resources, pp 191-212, Houston.
- Von Huene, R., Klaeschen, D., and Fruehn, J. (1999). Relation between the subducting plate and

- seismicity associated with the great 1964 Alaska earthquake. *Pure Appl. Geophys.* **154**, 575-591.
- Wahr, J., and Wyss, M. (1980). Interpretation of postseismic deformation with a viscoelastic relaxation model. *J. Geophys. Res.* **85**, 6471-6477.
- Wesson, R. L. Frankel, A. D., Mueller, C. S., and Harmsen, S. C. (1999). Probabilistic seismic hazard maps of Alaska, US Geological Survey Open File Report 99-36.
- Whitten, C. A. (1972). Introduction: An evaluation of the geodetic and photogrammetric surveys. In "The Great Alaska Earthquake of 1964, Seismology and Geodesy," National Academy of Sciences, Washington, D.C., pp. 415- 417.
- Williams, C.A., McCaffrey, R., and Qamar, A. (2001). Inversions of geodetic data for rotation pole and deformation mechanism at subduction zones: Applications to Cascadia, Sumatra, and Alaska. *EOS Trans. AGU* **82**.
- Wolf, L. W., and Davies, J. N. (1986). A comparison of acceleration values derived from recent seismic hazard studies of the Anchorage area, Alaska: Part I. Public-data File 86-85, Alaska Division of Geological and Geophysical Surveys, Fairbanks.
- Wolf, L. W., Stone, D. B., and Davies, J. N. (1991). Crustal structure of the active margin, south central Alaska: An interpretation of seismic refraction data from the Trans-Alaska crustal Transect. *J. Geophys. Res.*, **96**, 16455-16469.
- Wyss, M., and Brune, J. N. (1972). A complex multiple rupture, In "The Great Alaska Earthquake of 1964, Seismology and Geodesy," National Academy of Sciences, Washington, D.C., pp. 60- 64.
- Yeats, R. S., Sieh, K., and Allen, C. R., (1996). Chapter 8, Strike –Slip Faults. In "The Geology of Earthquakes", Oxford University Press, New York.

- Zhao, D. Christensen, D., and Puplan, H. (1995). Tomographic imaging of the Alaska subduction zone. *J. Geophys. Res.* **100**, 6487-6504.
- Zweck, C., Freymueller, J. T., and Cohen, S. C. (2002a). Three dimensional elastic dislocation modeling of the postseismic response to the 1964 Alaska Earthquake. *J. Geophys. Res.*, **107**, 10.129/2001JB000409.
- Zweck, C., Freymueller, J. T., and Cohen, S. C. (2002b). The 1964 Great Alaska Earthquake: Present day and cumulative postseismic deformation in the western Kenai Peninsula. *Phys. Earth Planet. Inter.*, **132**, 5-19.

Table 1: Uplift Rate (mm/yr) and Rate Change (mm/yr²) Estimates at Tide Gauge Sites in Southcentral Alaska

Tide Gauge Site	Rate: SP91 ¹	Rate: CF00 ²	Rate: LET02 ³	Rate Change: CF00 ²	Rate Change: LET02 ³
Anchorage	1.0 ± 2.2	2.7 ± 0.9	3.1 ± 0.3		0.48 ± 0.08
Cordova	-6.8 ± 1.1	-4.7 ± 1.0	-3.65 ± 0.12	0.39 ± 0.06	0.48 ± 0.04
Kodiak	20.4 ± 1.3	16.5 ± 0.7	16.4 ± 0.2	-0.67 ± 0.11	-0.66 ± 0.04
Nikiski	21.6 ± 2.0	11.9 ± 0.8;	15.6 ± 0.4		-0.80 ± 0.08
Seldovia	10.1 ± 1.8	11.3 ± 0.8	10.0 ± 0.2	0.16 ± 0.18	
Seward	2.8 ± 1.5	3.2 ± 0.7	2.7 ± 0.2	0.09 ± 0.15	0.54 ± 0.04
Valdez	-2.6 ± 1.4	2.2 ± 0.2	1.66 ± 0.14	0.76 ± 0.14	0.78 ± 0.04

¹SP91=Savage and Plafker (1991) used an eustatic sea level rise of 2.4 ± 0.9 mm/yr and assumed the effect of postglacial rebound to be 0.5 ± 0.5 mm/yr.

²CF00=Cohen and Freymueller (2000) assumed that the combined effect of eustatic sea level rise and postglacial rebound is 2 mm/yr.

³LET02=Larsen *et al.* (2002) used a eustatic sea level rise of 1.09 ± 0.21 mm/yr.

Table 2: Model Parameters for Eastern Kenai Deformation Models of Freymueller *et al.* (2000).

Model	dip ₁ (deg)	width ₁ (km)	α_1	ext (km)	dip ₂ (deg)	width ₂ (km)	α_2	χ_v^2
1 locked interface plane with dip angle fixed	3.5	297.07	-1.0					2.919
1 locked interface plane with dip angle estimated	1.998	272.46	-1.0					2.318
1 locked interface plane and 1 postseismic slip plane with rake estimated	5.518	273.67	-0.972	43.36	44.94	399.80	1.21ds 0.19ss	1.978

The parameter, dip₁, is the dip angle of the locked plate interface, width₁ is the along-dip width of the locked portion of the plate interface, ext is the width of the gap between the locked plate interface and a postseismically creeping portion of the plate interface, dip₂ is the dip angle of the postseismic creep plane, and χ_v^2 is the reduced chi-squared fit between the model and observation velocities. In the third model, ds indicates the dip slip component and ss the strike-slip component. The parameters α are multiplicative factors for the plate velocity that describe the inferred steady state slip. Negative values indicate back-slip (a proxy for a fully, $\alpha = 1$, or partially, $-1 < \alpha < 0$, locked plate interface). A value of $\alpha = 0$ indicates no slip, or an uncoupled plate boundary, and a positive value for α indicates forward slip due, for example, to postseismic creep or episodic aseismic slip. The subscripts on α is analogous to that for dip and width.

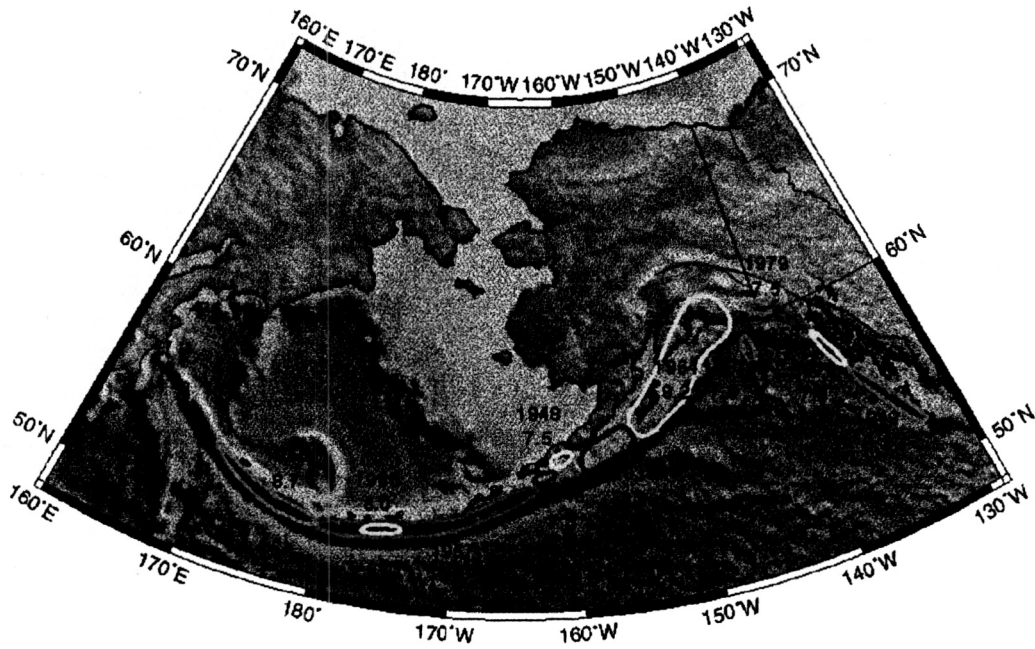


Fig. 1. Rupture areas of large historic earthquakes in Alaska. From Wesson *et al.* (1993) and Plafker *et al.* (1994a).

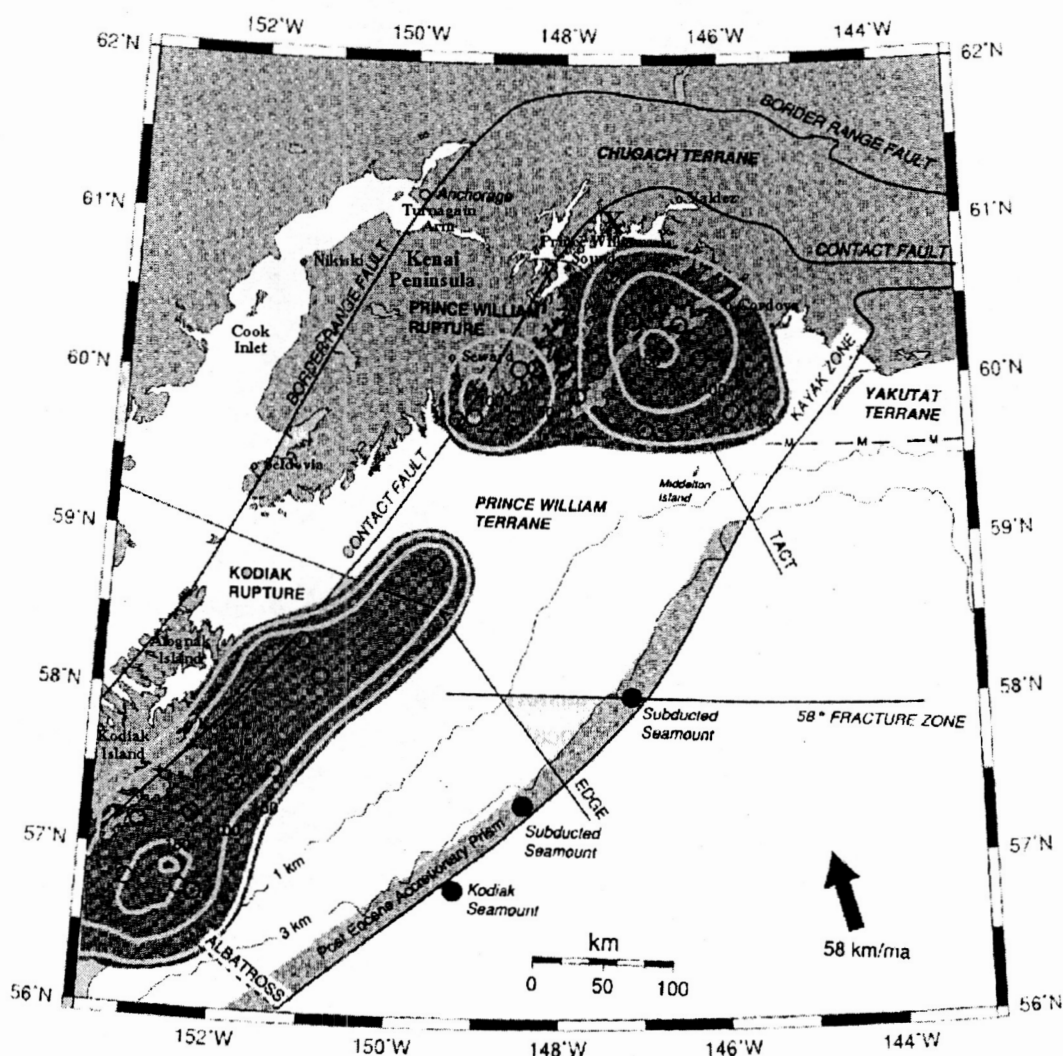


Fig. 2. Contemporary tectonic setting and features. Among the features shown are the relative velocity vector between the Pacific and North America Plates, the locations of the Prince William and Kodiak asperities (defined by intense aftershock occurrence with the aftershock strain energy being indicated by contours of the number of equivalent $M = 3$ earthquakes), the Chugach, Prince William and Yakutat Terranes, and the Border Ranges and Contact Faults. The Peninsular Terrane lies north of the Border Ranges Fault. The open circles are the locations of aftershocks that occurred within the first five hours of the main shock and had $M > 6$. Also shown are the TACT, EDGE, and ALBATROSS seismic reflection and refraction lines and several cities mentioned in the text. Modified from Von Huene *et al.* (1999).

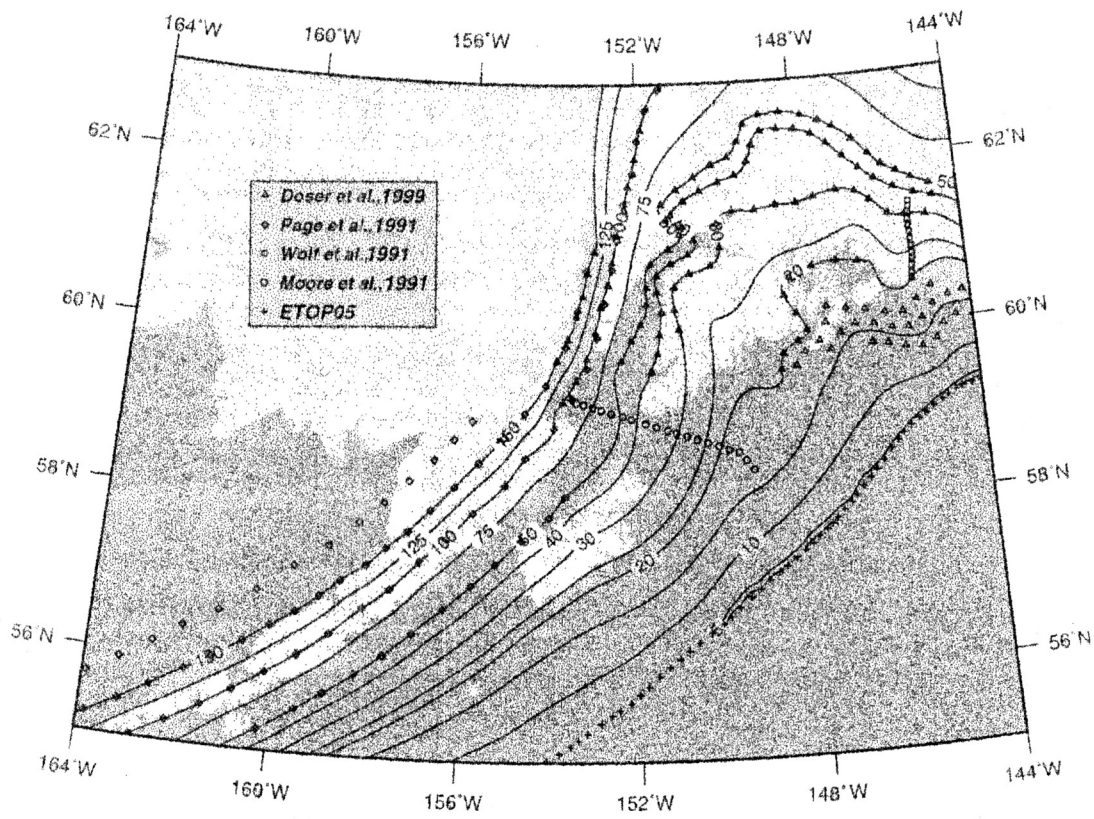


Fig. 3 Contours of plate interface depth (in km). The original data sources are indicated. From Zweck *et al.* (2002a).

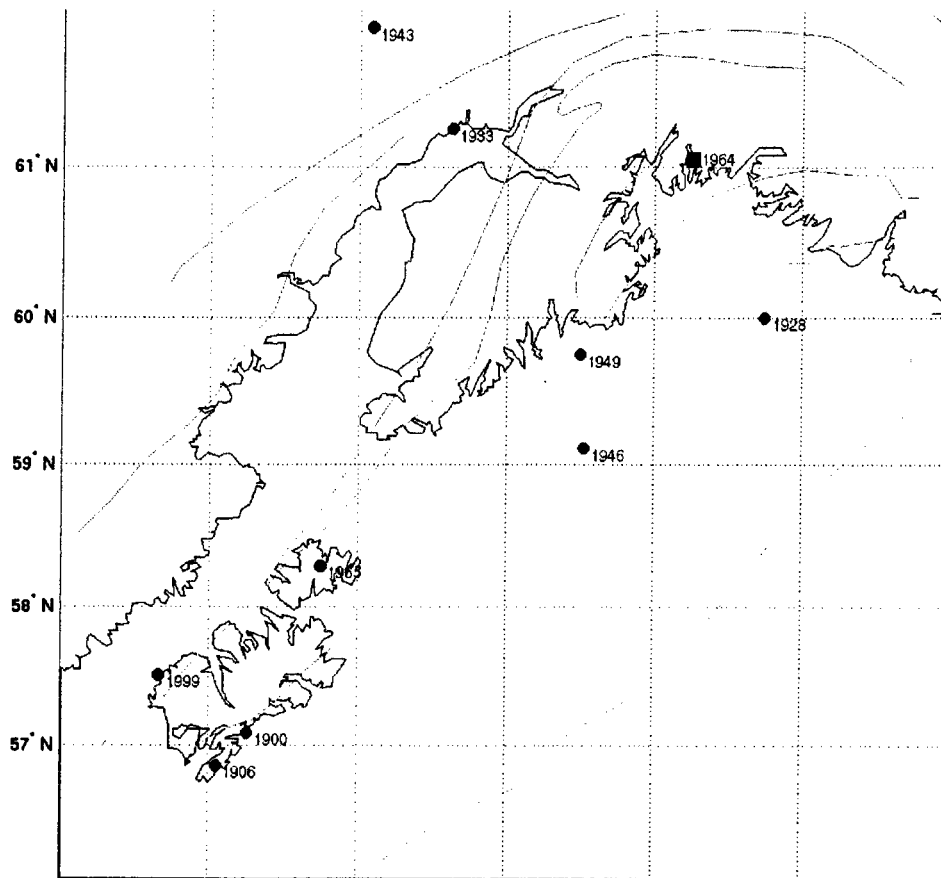


Fig. 4. Large (magnitude 7 or greater), shallow (focal depth less than 60 km) earthquakes in southcentral Alaska since 1900 from the catalog of the Alaska Earthquake Information Center. The epicenter of the 1964 earthquake is shown as a black box. Earthquakes that occurred prior to 1964 are shown as filled red circles, those occurring since 1964 are shown as filled blue circles.

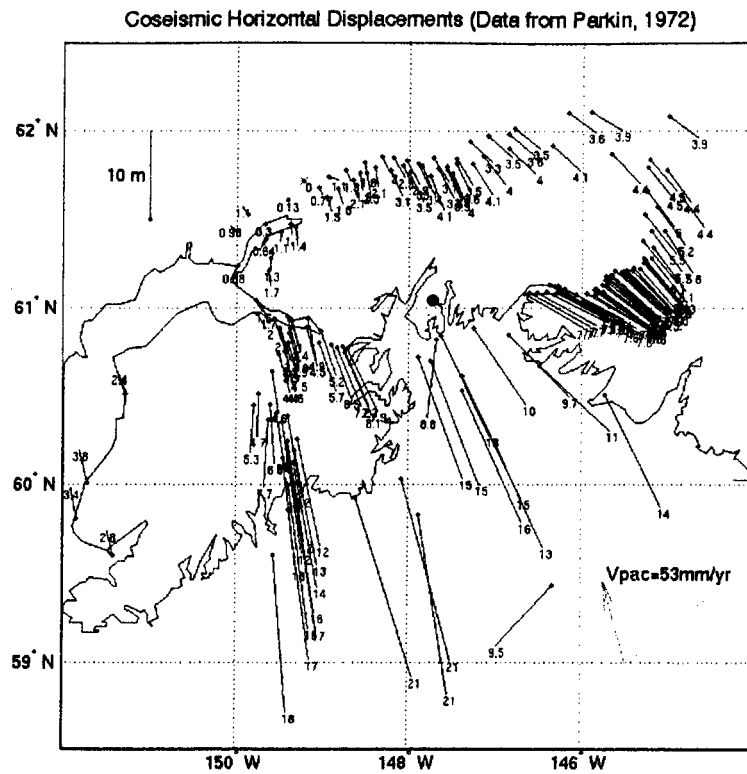


Fig. 5. Coseismic horizontal displacements for the 1964 Prince William Sound earthquake. The displacements are relative to site Fishhook, located at $61^{\circ}43.0' \text{ N}$, $149^{\circ}14.0' \text{ W}$. The data are from Parkin (1972).

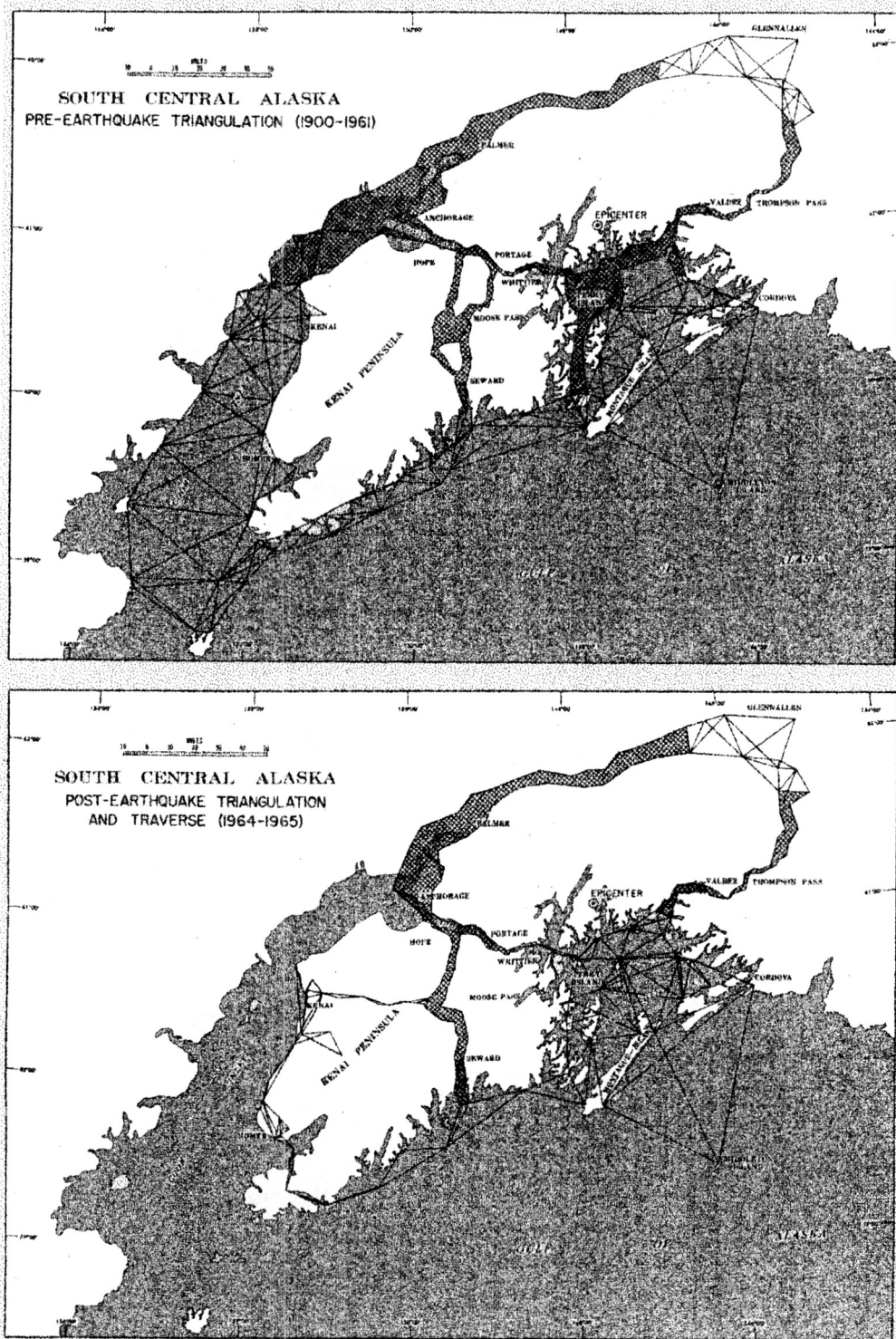


Fig. 6. Portions of the geodetic networks used in deduce coseismic displacements and strains changes. The top panel shows the pre-earthquake network for southcentral Alaska; the bottom panel shows the post-earthquake network for the same region. From Parkin (1972).

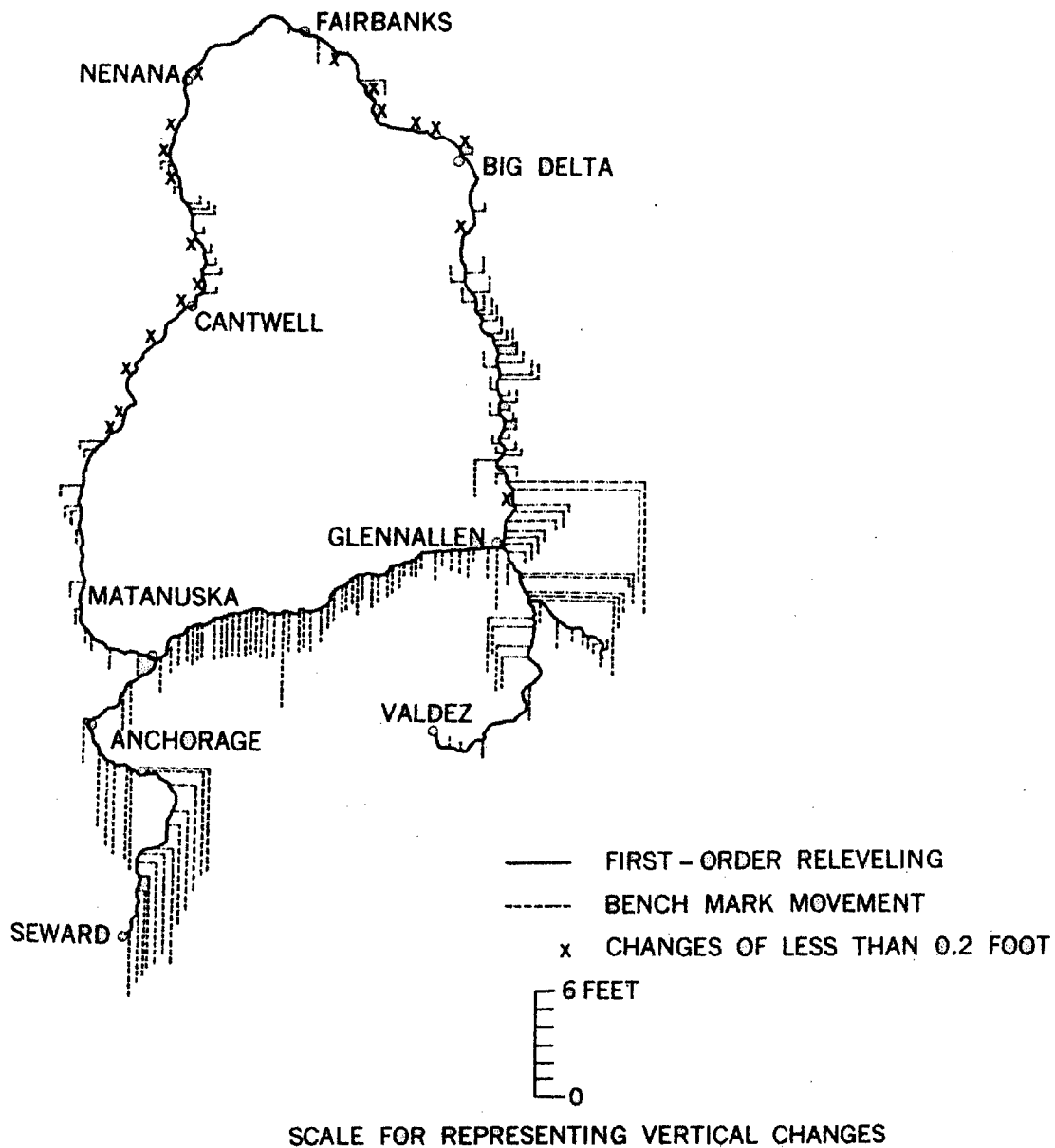


Fig. 7. Coseismic vertical displacements from leveling observations. The leveling segments and the dates of the pre-earthquake and post-earthquake surveys are as follows: Seward to Matanuska: 1922/23, 1964; Matanuska to north of Glennallen: 1944, 1964; Glennallen to Valdez: 1923, 1964; North of Glennallen to Fairbanks: 1923, 1964; Matanuska to Fairbanks: 1922, 1965. From Small and Wharton (1972).

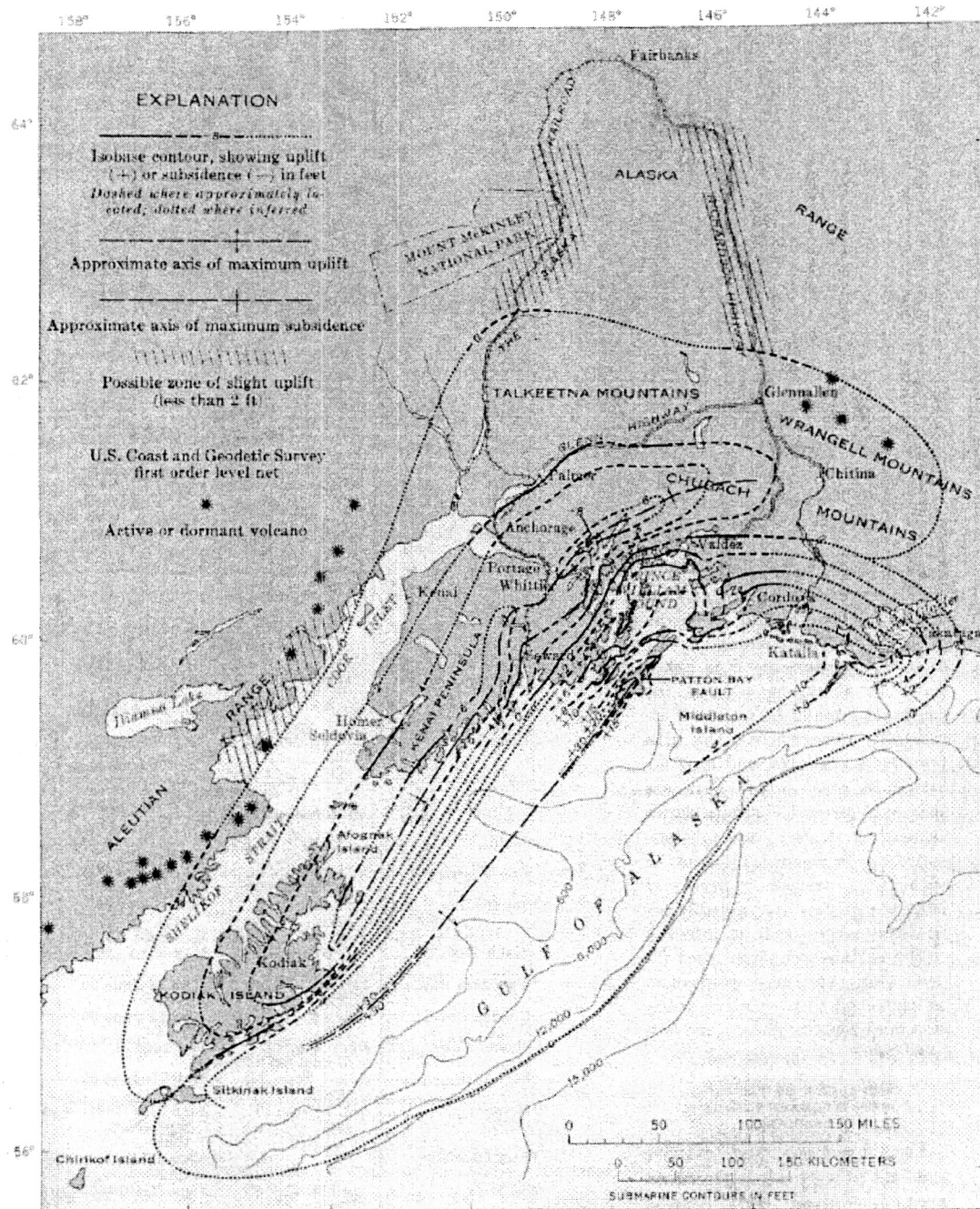


Fig. 8 Coseismic uplift contours (in feet) as deduced from both geodetic and geologic information. From Plafker (1972).

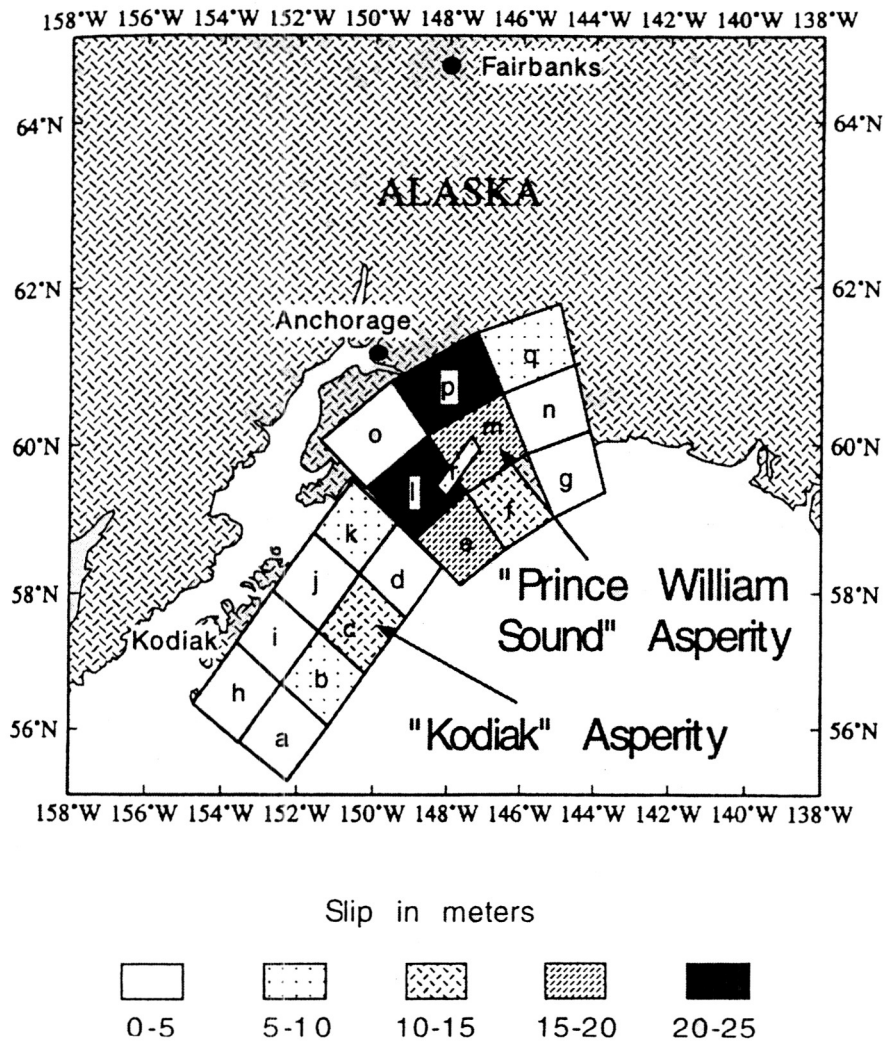


Fig. 9. Coseismic slip distribution as deduced from an elastic dislocation model using geodetic and tsunami data. From Johnson *et al.* (1996).

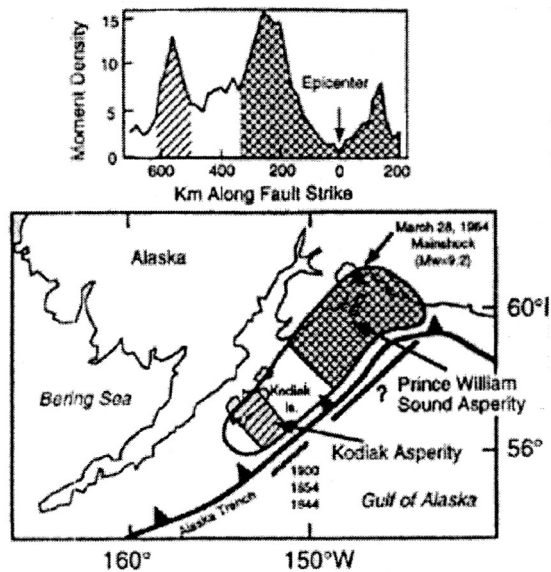


Fig. 10. Asperity distribution as determined by Christensen and Beck (1964) and presented in Johnson *et al.* (1996). The upper figure shows the along-strike moment density release in units of 10^{26} dyne-cm/km. The lower figure shows the map view as the asperities. The dates of historical earthquakes in the Kodiak segment are listed.

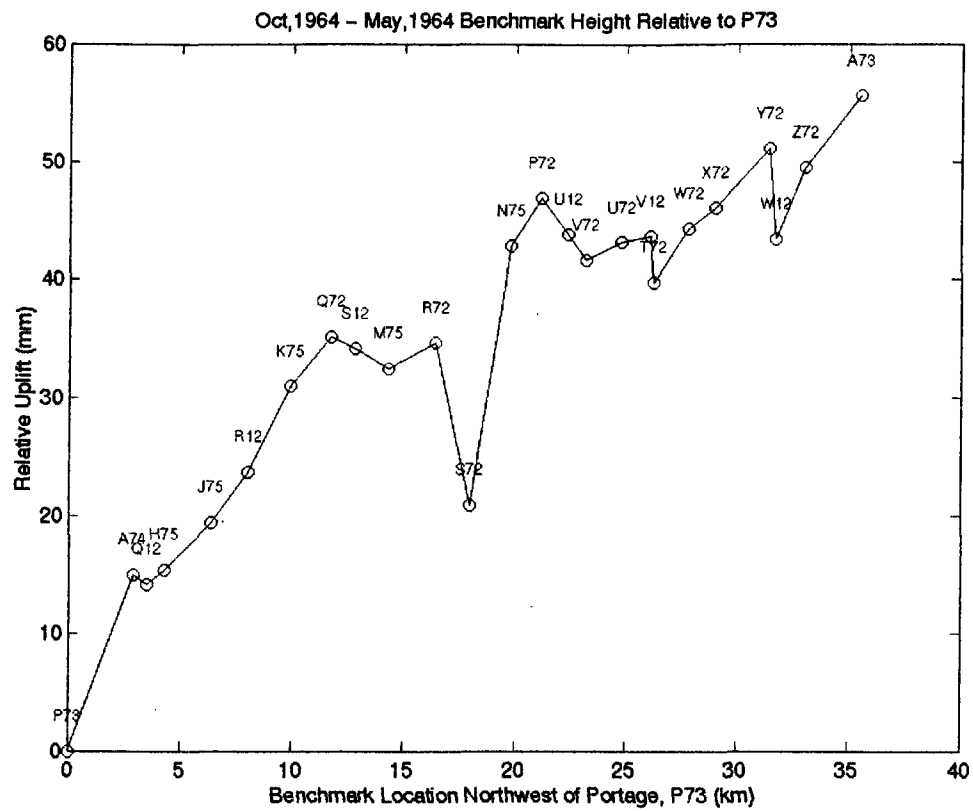


Fig. 11. Postseismic uplift relative to site P73, near Portage, as determined from leveling differences deduced from surveys along Turnagain Arm in May and October, 1964. Portage is located near the southern end of Turnagain Arm (Fig. 2). From Small and Wharton (1972).

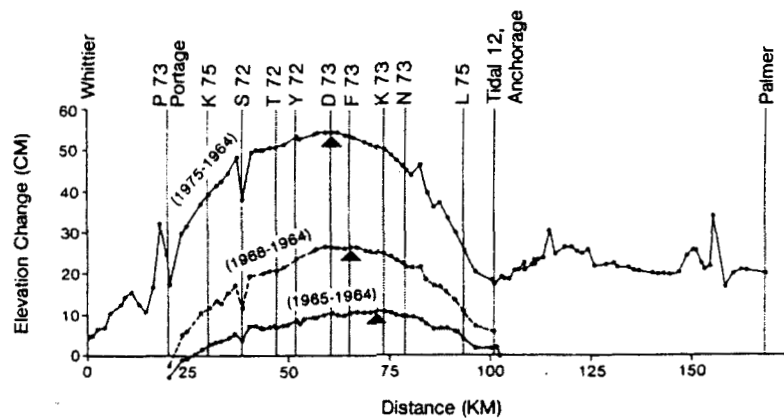


Fig. 12. Postseismic uplift derived from repeated leveling along a route beginning at Whittier and extending northwest along Turnagain Arm between Portage and Anchorage, then mostly northeast toward Palmer. The absolute values of the uplift may be biased. They are based on the assumption that Anchorage was moving upward at a constant rate of 16.3 mm/yr. (see text for discussion). From *Brown et al* (1977).

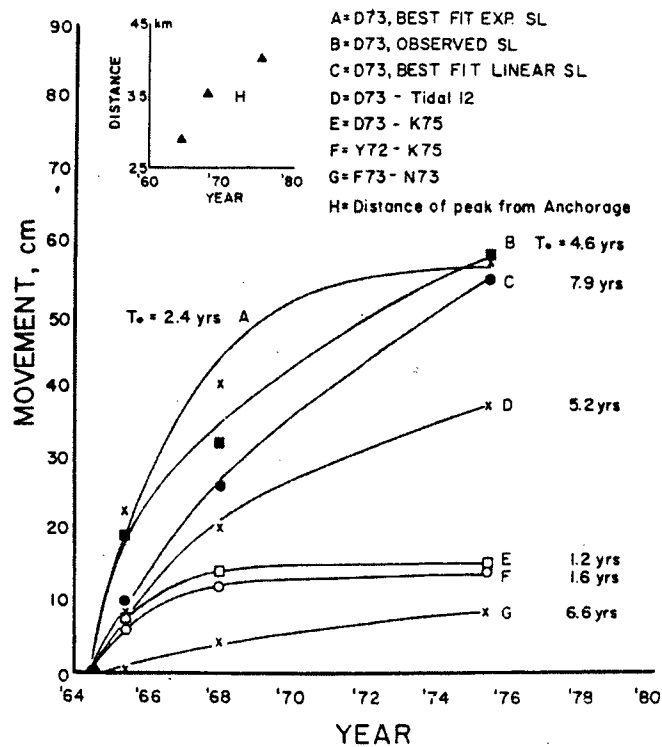


Fig. 13. Time dependence of postseismic uplift as deduced from leveling data along Turnagain Arm. In each case the data are fit (in a least squares sense) to an exponential function with time constant, T_0 . Curves A, B, and C refer to the motion of site D73 in Fig. 12. In A, an exponential uplift function is also assumed for the sea level tide gauge site in Anchorage. In B, the observed positions of the tide gauge are used. In C, a linear, i.e. constant rate, uplift is assumed for the tide gauge. Curves D, E, F, and G are relative uplifts assuming that the uplifts of the points being considered have the same time constant. The inset shows the migration of the point of maximum uplift in Fig. 12. From Brown *et al.* (1977).

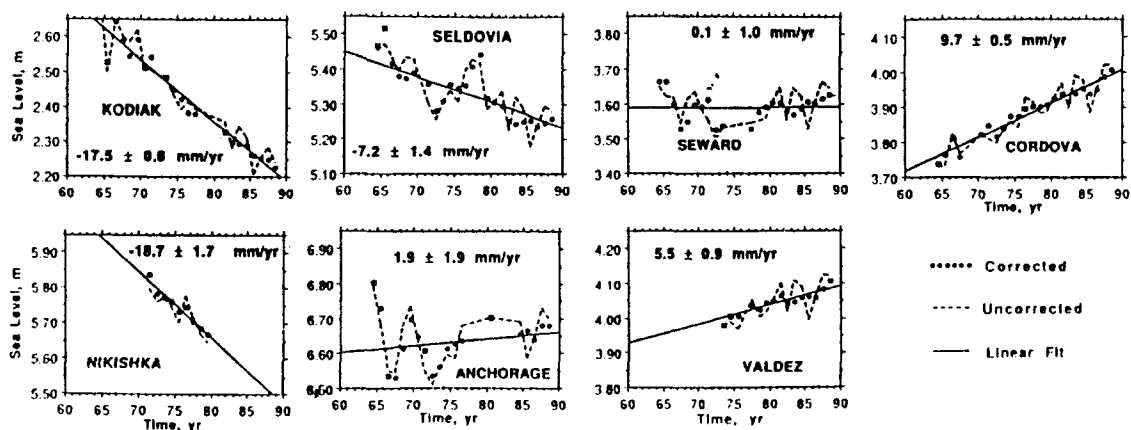


Fig. 14. Apparent sea level heights deduced from southcentral Alaska tide gauges. The data corrected for short-period fluctuations in sea level are shown as solid points. The solid lines show the least squares fit to the corrected data and the dashed lines connect the uncorrected data. The rates deduced from the least squares analysis are also shown. From Savage and Plafker (1991).

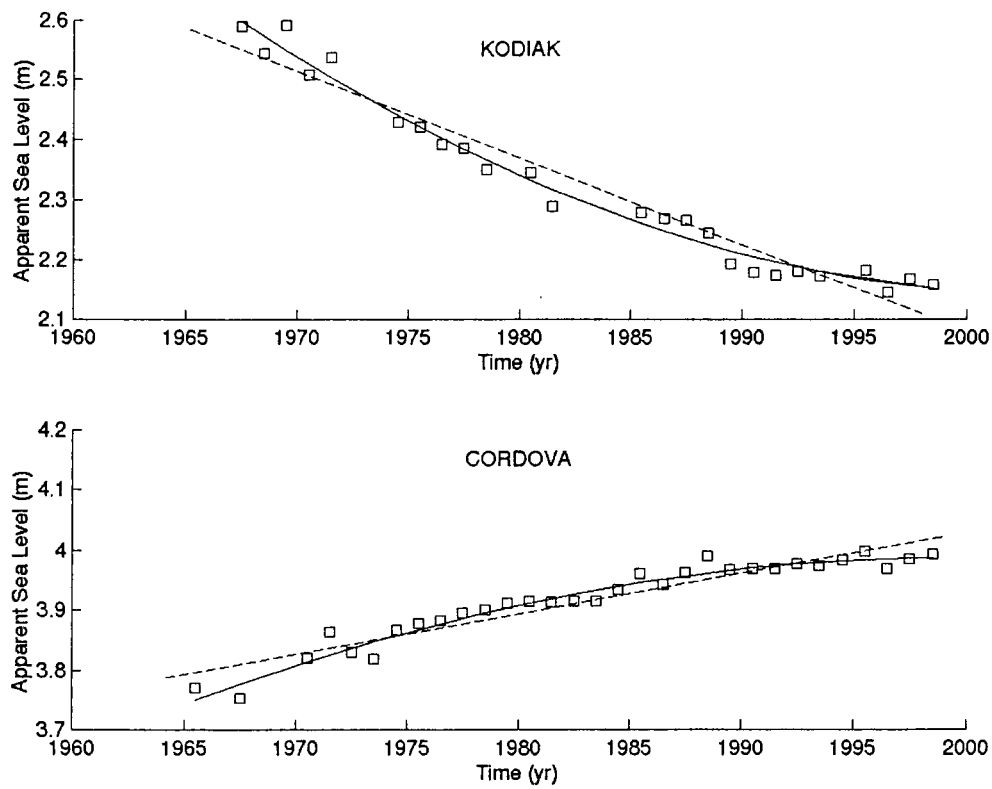


Fig. 15. Apparent sea level heights for the Kodiak and Cordova tide gauge sites. The corrected annual mean data are shown as open squares, the line derived from a linear least squares analysis by dashes and the curve derived from a quadratic least squares analysis by the solid line. From Cohen and Freymueller (2001).

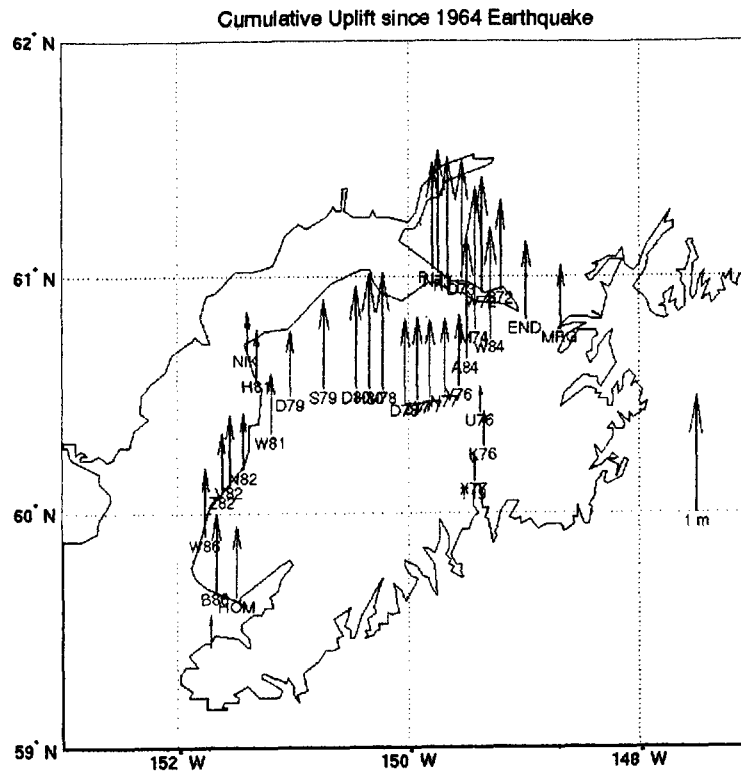


Fig. 16. Uplift since 1964 on Kenai Peninsula and adjacent areas. The uplift is plotted relative to tide gauge station T19 at Seward. The red arrows show the uplift as determined from tide gauge observations, the blue arrows show the uplift determined by comparing post-earthquake leveling height observations to contemporary GPS vertical position determinations (using relative geoid heights from the GEOID99 model to transform between the orthometric and geometric coordinate systems). Although there are both random and systematic measurement errors in both the leveling and the GPS measurements, the uncertainty in the deduced uplift is dominated by the poorly known baseline-length-dependent error in the relative geoid height. Thus, the uplift of sites close to Seward are better constrained than those farther away.

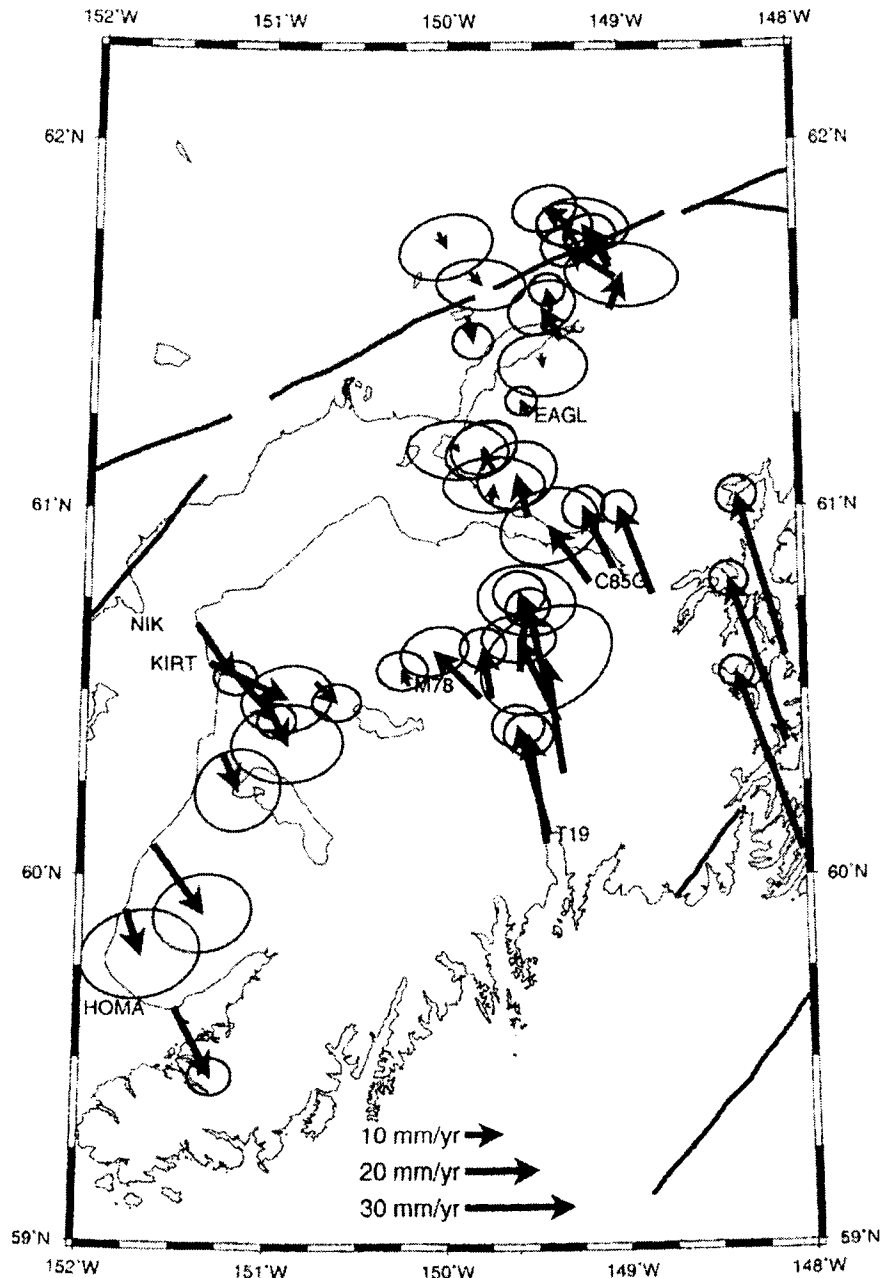


Fig. 17. Horizontal crustal velocities as deduced from GPS observations made in the middle and late 1990's on the Kenai Peninsula and surrounding areas (The figure is from Freymueller *et al.* (2000) and includes data from their field work and from that of Savage *et al.* (1998)).

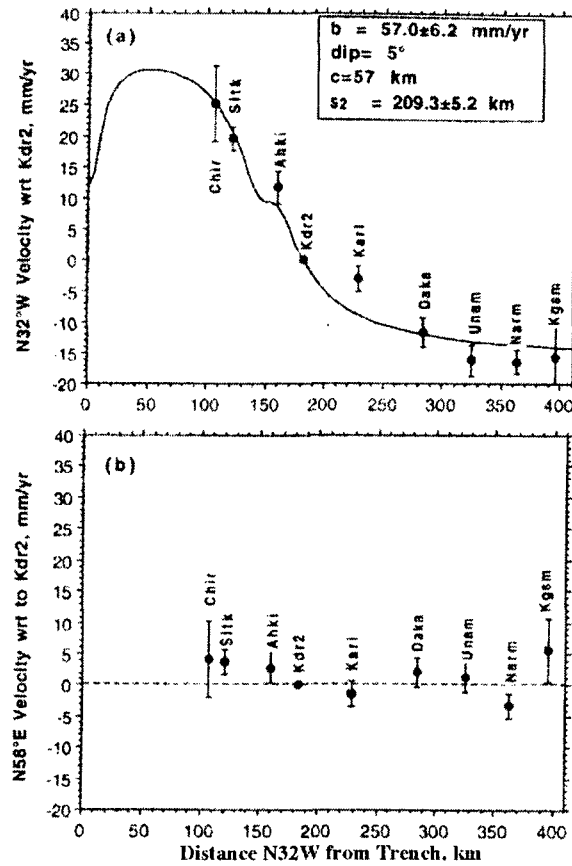


Fig. 18. Velocity profiles in the Katmai-Kodiak Island region relative to site Kdr2 (on Kodiak Island) as a function of the distance N32°W from the trench. The error bars are two standard deviations. The upper panel shows the trench normal velocities (strictly speaking the velocities in the N32°W direction), while the lower panel shows the trench parallel velocities. The curve in the upper panel is the computed relative displacements using elastic dislocation theory with the following parameters: dip angle -5° , horizontal distance from the trench to the point above the up-dip end of the locked plate interface, c , -57 km, length of locked plate interface, s_2 , 209.3 km. Although the sites are shown as a function of distance from the trench there are considerable along-strike variations in their locations as well. Chir is on Chirikof Island, well southwest of Kodiak Island (KI), Sitk is on Sitkinak Island, slightly southwest of KI, Ahkl and Kdr2 are near the trenchward shore of KI, Karl is on the far shore of KI, and Daka, Unam, Narm, and Kgsm are on the Alaska Peninsula, across Shelikof Strait from KI. From Savage *et al.* (1999).

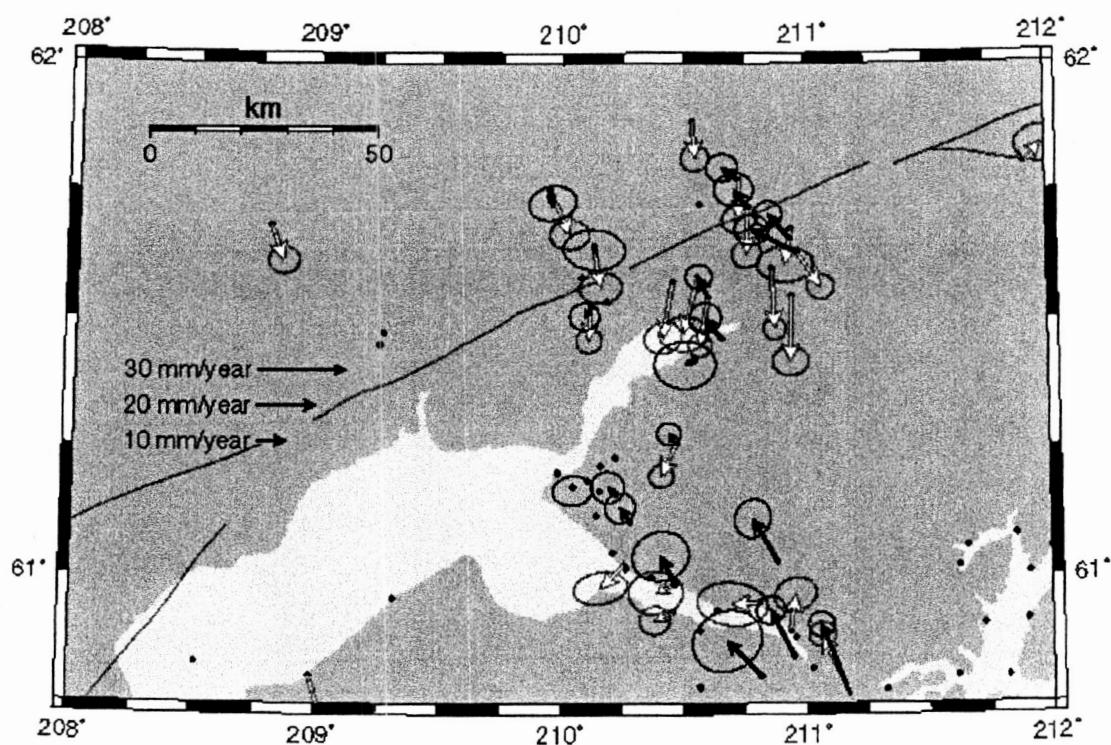


Fig. 19. Change in velocities near Anchorage as reported by Freymueller *et al.* (2001) The solid black vectors show GPS-derived velocities near Anchorage before 1998, the open red vectors show velocities from 1998 to 2001.

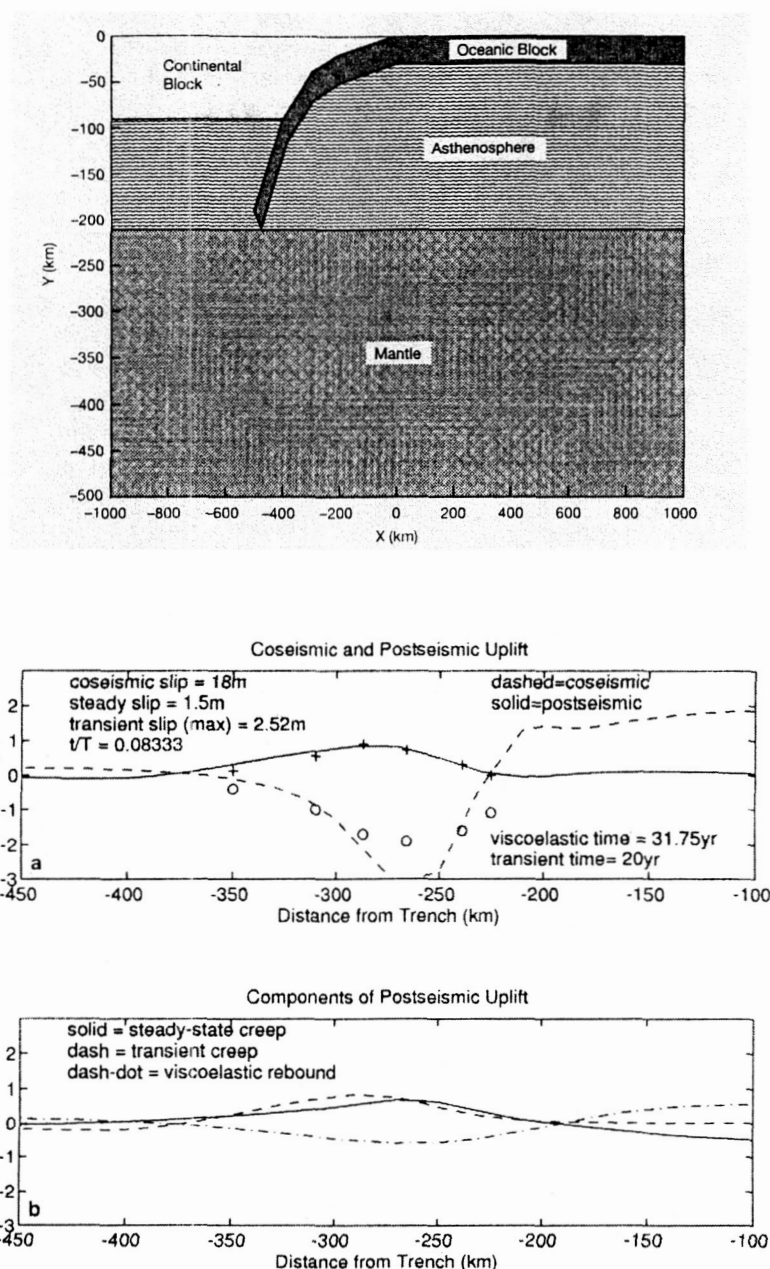


Fig. 20. (Top panel) A schematic representation of a viscoelastic model used in finite element calculations of crustal deformation. The subducting and overthrust plates are elastic; the asthenosphere is viscoelastic. For the calculations in this figure the asthenosphere extends to great depth, but in other calculations it is underlain by an elastic or higher viscosity mantle. (Middle panel) Cumulative postseismic uplift (solid line) on the Kenai Peninsula from 1964 to 1993 predicted by a combined deep transient creep - viscoelastic rebound model. The dashed line shows the predicted coseismic subsidence while the plus signs and circles give the measured uplift reported in Cohen *et al.* (1995) and the coseismic subsidence from the model of Holdahl and Sauber (1994), respectively. (Bottom panel) Contributions to the predicted postseismic uplift from steady plate convergence, viscoelastic rebound, and transient creep. From Cohen (1996).

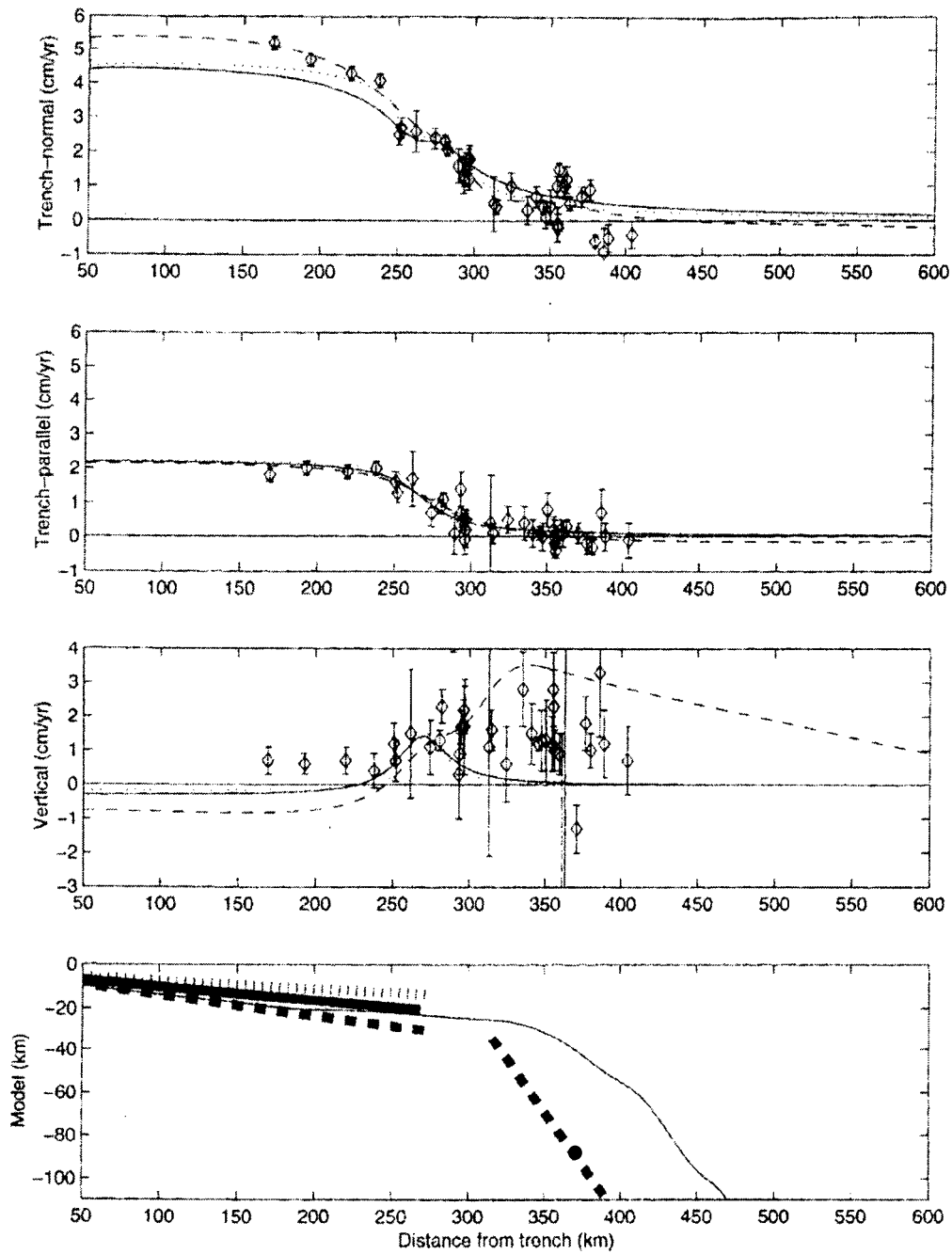


Fig. 21. Elastic dislocation models for eastern Kenai Peninsula. The solid curve is the best fitting model with dip constrained to match the inferred slab position, the dotted curve is the best fitting model with the dip angle estimated, and the dashed model is the best fitting model that includes both strain accumulation and postseismic creep-at-depth. For the latter model, the dip angle of the shallow plane was fixed. The parameters for these three models are shown in Table 2. From Freymueller *et al.* (2000).

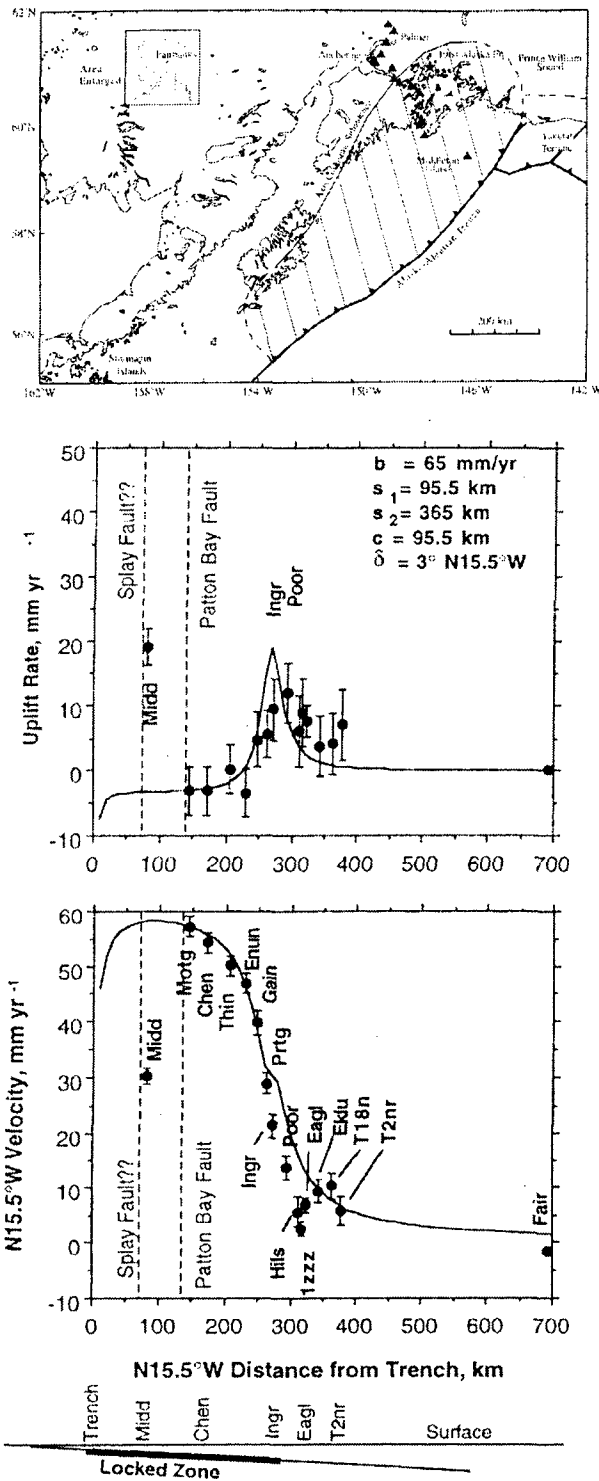


Fig. 22. The top panel shows a network of campaign GPS sites in southcentral Alaska. The horizontal and vertical velocities derived from observations at these sites are shown in the middle panels while the bottom panel shows a schematic sketch of the plate interface geometry used in an elastic dislocation calculation (whose results are shown as solid lines in the data panels). From Savage *et al.* (1998).

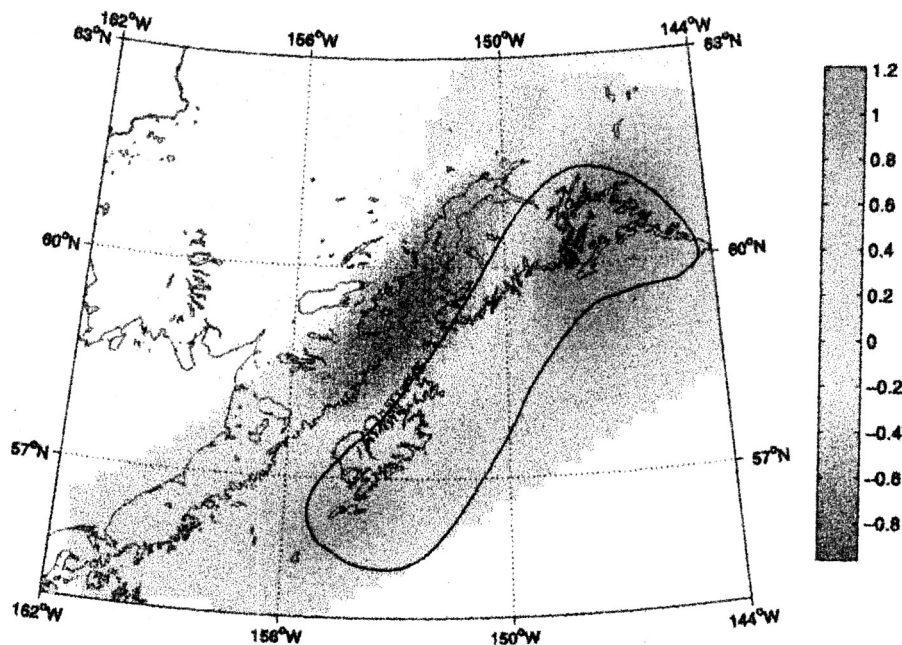


Fig. 23. Spatial variation in plate coupling as derived from inversion of an elastic model. The color coded parameter, a , is defined by setting the slip on the plate interface to be $(1-a)V$ where V is the relative plate velocity. Thus a is defined somewhat differently than α in Table 2. When $a = 1$, there is full coupling; when $a = 0$, the plates are uncoupled; and when a is negative, the surface velocities are trenchward. From Zweck *et al.* (2000a).

Crustal Deformation in Southcentral Alaska: The 1964 Prince William Sound Earthquake Subduction Zone

Steven C. Cohen (Geodynamics Branch, NASA Goddard Space Flight Center) and
Jeffrey T. Freymueller (Geophysical Institute, University of Alaska Fairbanks)

Popular Summary

This article, for *Advances in Geophysics*, is a summary of crustal deformation studies in southcentral Alaska. In 1964, southcentral Alaska was struck by the largest earthquake (moment magnitude 9.2) occurring in historical times in North America and the second largest earthquake occurring in the world during the past century. Conventional and space-based geodetic measurements have revealed a complex temporal-spatial pattern of crustal movement. Numerical models suggest that ongoing convergence between the North America and Pacific Plates, viscoelastic rebound, aseismic creep along the tectonic plate interface, and variable plate coupling all play important roles in controlling both the surface and subsurface movements. The geodetic data sets include tide-gauge observations that in some cases provide records back to the decades preceding the earthquake, leveling data that span a few decades around the earthquake, VLBI data from the late 1980's, and GPS data since the mid-1990s. Geologic data provide additional estimates of vertical movements and a chronology of large seismic events. Some of the important features that are revealed by the ensemble of studies that are reviewed in this paper include: (1) Crustal uplift in the region that subsided by up to 2 m at the time of the earthquake is as much as 1 m since the earthquake. In the Turnagain Arm and Kenai Peninsula regions of southcentral Alaska, uplift rates in the immediate aftermath of the earthquake reached 150 mm/yr, but this rapid uplift decayed rapidly after the first few years following the earthquake. (2) At some other locales, notably those away from the middle of the coseismic rupture zone, postseismic uplift rates were initially slower but the rates decay over a longer time interval. At Kodiak Island, for example, the uplift rates have been decreasing at a rate of about 7 mm/yr per decade. At yet other locations, the uplift rates have shown little time dependence so far, but are thought not to be sustainable throughout the several hundred year recurrence time for great earthquake. The nearly 10 mm/yr uplift rate at Seldovia on the Kenai Peninsula is an example.

Horizontal crustal movements are particularly interesting. On the eastern side of the Kenai Peninsula the surface movements are directed north-northwest, consistent with the accumulation of stress at the locked tectonic plate interface. However, on the western side of the Peninsula the surface movements are oriented south to south-southeast. These enigmatic motions of the western side of the Peninsula are inconsistent with stress accumulation and appear to be due to a long term or delayed transient response to the 1964 earthquake. The spatial variability in the contemporary crustal movements mirrors in some respects the complex pattern of crustal movement of the earthquake: large crustal movements occurred near Prince William Sound and the eastern Kenai region, secondary movement occurred near Kodiak Island, but relatively little motion occurred in-between. For a few years, starting in 1998, the direction of crustal movement near Anchorage reversed direction. This motion appears to be associated with a sustained, aseismic deep slip event ("slow earthquake") in the region.

No single deformation mechanism can account for all the features that have been observed. Viable models must include elastic and viscoelastic rheologies and time-dependent interface friction. The models must also take into account the fact the eastern portion of the Pacific Plate is overlain by a fragment of continental material, known as the Yakutat terrane, and the fact that surface faults, in addition to the plate interface, can locally accommodate significant amounts of the plate convergence.